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The Last Stand of the Great American Carbonate Bank: Tectonic Activation of the Upper Ordovician Passive Margin in Eastern North America.

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ABSTRACT

The Upper Ordovician (450-460 Ma) Chazy, Black River and Trenton groups of eastern North America record the tectonic activation of a passive carbonate platform. Associated with tectonic collision and development of a peripheral-type foreland basin, the tectonic history is complicated by the occurrence of two basin-forming episodes. Although considered analogous by previous authors, a growing list of observations indicate that the stratigraphic record from both foreland basins and their coeval Great American Carbonate Bank (GACB) contain somewhat different, non-analogous, signatures. A number of key issues arise when comparing both tectonic episodes. These include: 1) non-analogous spatial-temporal scales whereby the northern (Vermontian) tectophase is much larger and of longer duration compared to the southern (Blountian) tectophase; 2) sedimentary provenance analyses show a more mafic contribution in the northern basin fill compared to that of the southern basin; 3) the position of K-bentonite swarms relative to basin filling phases is non-analogous; and 4) the location and timing of Ordovician volcanism/plutonism shows a pronounced change after the first tectophase.

Thus, important research questions for this study include: A) How does the architecture of the foreland basin complex and adjacent GACB change spatially and temporally during each distinct tectophase?; B) Can provenance differences between tectophases be explained relative to tectonic events in the orogen?; C) What is the timing and significance of K-bentonite position and timing of plutonism/volcanism relative to foreland basin fill episodes?; 4) What inferences are gained from theoretical modeling of load geometries and foreland basin evolution when considered with empirical data from the Taconic Orogeny; and 5) Is a new model for the Taconic Orogeny needed to explain the growing list of incongruities?

In order to investigate these questions, a refined, high-resolution, sequence stratigraphic framework has been constructed and utilized to re-calibrate Upper Ordovician strata during the last stand of the GACB. The refined framework was constructed using the occurrence of sequence boundaries, erosional surfaces, transgressive, highstand, and regressive systems tracts, hardgrounds and flooding surfaces, laterally extensive and unique lithofacies, highly repetitive sub-meter scale cyclic intervals, updated macrofaunal and microfaunal biostratigraphies, as well as known chemostratigraphic and event stratigraphic data (Nd, C, & Sr isotopic excursions, K-bentonite horizons, seismite horizons, etc.). Also defined for the first time are a number of time-restricted facies including widespread siliciclastic events, chert-rich intervals, and pronounced calcification events. As such, an improved temporal and spatial chronology of thirteen time-slices (~ 1 million year duration) has been produced for the Ashbyan, Mohawkian, to earliest Cincinnatian interval.

This integrated stratigraphic model is used to link the sedimentary record of the foreland basin complex to climatic changes, sea-level oscillation, and specific tectonic events in the orogen. These data provide an understanding of the evolution of the foreland basin and its sub-components including the backbulge, forebulge, and foredeep basins. Moreover, coupled with newly calibrated strontium isotopic curves, this study has allowed for the development of a new model for the Taconic Orogeny that helps to explain differences in the timing and spatial relationships between the southern and northern tectophases.

ACKNOWLEDGEMENTS

This lengthy dissertation is the result of a long-term process and it is hard to remember exactly when it started, but one thing that cannot be forgotten is the support that I have received from start to finish. This dissertation is dedicated to my many advisors, colleagues, friends, and family – especially my wife Angel, and my children Hannah, Jenna and Ethan. Throughout the entirety of our married life, my Angel has been my rock. Steadfast, solid, and un-weatherable, without her support and love this would not have been possible. My children, born in the middle of this process, have always known me to be “a student.” Although this specific journey is now finished, I hope they will continue to see “Dr. Daddy,” as a student of this fascinating and rewarding discipline. I hope that, in time, they too will come to realize the joys and benefits of education, and that they will tackle their own Himalayan-scale journey.

Having arrived in the UC Department of Geology in the fall of 1998, the past decade has been a roller coaster and I cannot fathom this experience in the company of any other group of dedicated people. I attribute much of my intellectual and professional growth to the inspiring faculty and staff of the department and to the many students from whom I have learned much. Although there are many, Alex Bartholomew, Patrick McLaughlin, Susannah Taha-McLaughlin, and Michael Desantis are among my most cherished graduate student colleagues. Through their humor, candor, and their own personal journey’s I have gained much and continue to do so. Kudos and thanks are due to Sandi Cannell, Ginny Chasteen, Alice McDade, Mike Menard, and Evelyn Pence. As staff of the Department of Geology, they kept our computers and research equipment running, our stomachs full, our stipends coming, and helped support us in more ways than we can count.

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INTRODUCTION

The onset of major orogenic activity along continental margins on the scale of many millions of years has been documented to have profound effects on the adjacent craton, its depositional environments, and therefore its biotas (Ulrich. 1911a; 1911b; Grabau. 1940; Sloss. 1963; Rodgers, 1971; Dennison, 1994; Ettensohn, 1994; Bezusko, 2001; Harper, 2002). However, specific details regarding the scale and timing of basin modification associated with tectonic activity are often poorly resolved — usually only scaled to resolutions of a few million years at best in the Paleozoic.

Given what is known from analogous modern tectonic settings (Londono & Lorenzo, 2003, 2004; Hamson, 2005) where rates of tectonic convergence and associated uplift and subsidence are relatively high – "up to 4 times higher than the rate of eustatic fluctuation (Londono & Lorenzo, 2004)," the limited resolution of a few million years is inadequate to constrain the timing and degree of tectonic impact on local and basin-wide deposition. As a result there is debate as to what factors (allocyclic or autocyclic processes) are ultimately responsible for impacting deposition in these settings (Dennison, 1994; Ettensohn, 1991; Ettensohn, 1994). Consequently the stratigraphic expression of depositional sequences and associated patterns of faunal change (i.e. immigration, emigration, turnover, radiation, etc.) prior to and during orogenesis remain poorly resolved. Fortunately, progress is now being made especially in the Devonian where sequence stratigraphic and high-resolution biostratigraphic studies are being integrated in order to evaluate the onset of the Acadian Orogeny and its various tectophases (Ettensohn, 1987; Ver Straeten & Brett, 2000; Ver Straeten, 2002; Bartholomew & Brett, 2007; DeSantis. 2004).

The lack of a high-resolution chronology for evaluating stratigraphic change is especially true for the Upper Ordovician, both prior to and during the Taconic Orogeny. The sedimentary signatures from the rocks of the Black River and Trenton groups (Mohawkian Stage) indicate that Laurentia witnessed substantial topographic changes and associated oceanographic, and climatic alterations in a relatively short interval of time (<2 million years).

Such changes have been attributed to the collision of an island arc terrane with the eastern edge of proto-North America. This collision led to tectonic loading of the continental margin with subsequent subsidence of continental lithosphere into a flexural trough. Located along the margin of the former shallow shelf, the trough developed on top of and adjacent to epeiric sea environments that were pervasive during the deposition of the Black River Group (**figure 1**) (Fisher, 1962, 1977; Shanmugam and Lash, 1982; Quinlan & Beaumont, 1984; Diecchio, 1991; Ettensohn, 1991; Lehmann et al., 1994).

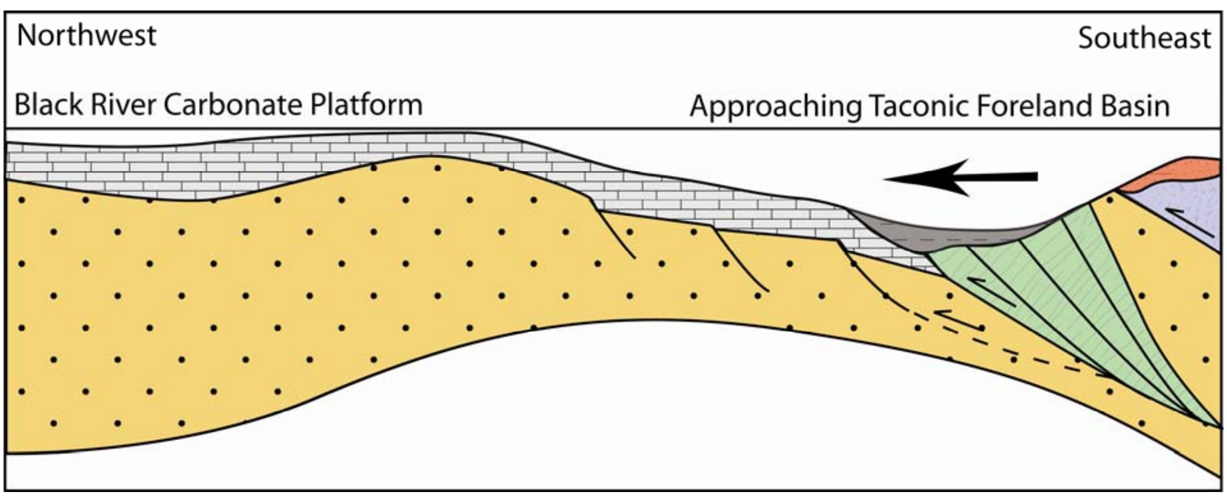


Figure 1: General model for the onset of the Taconic Orogeny and propagation of the Taconic Foreland Basin from southeast to northwest over a southeast-dipping subduction zone. Modified after Shanmugam and Lash, (1982)

Inboard of the continental margin, previous authors have noted the presence of widespread unconformities bracketing the Black River and Trenton groups (Keith 1988;

Ettensohn, 1991; 1994) as well as the presence of other unconformities in subjacent and overlying units such as the base Chazy disconformity. This is also equivalent to the Knox Unconformity of Sloss (1963) (Jacobi, 1981). These authors have interpreted these erosional unconformities as the result of lithospheric flexure and uplift associated with, but distant from, the onset and relaxation of plate margin tectonism (Beaumont, 1981; Jacobi, 1981; Walker et al., 1983; Quinlan & Beaumont, 1984; Ettensohn, 1991).

Although these long-standing interpretations have some empirical and theoretical support, they have been challenged by local and regional stratigraphic observations regarding the exact extent of unconformities, and the apparent lack of decoupling of sedimentary change in the foreland with sedimentary transitions on the craton as might be expected (Landing, 1988). The demise of the Great American Carbonate Bank (GACB) is thus coincident and correlated with the onset of the Taconic Orogeny. Nonetheless, the correlation between Taconic tectonism and the many physical and biotic events recorded during the transition into the active tectonic regime is still poorly resolved, despite a growing number of research publications describing these changes on the scale of local outcrop regions.

For this interval, the assembly of a highly resolved chronostratigraphy has historically been hindered by: 1) the lack of extensive lateral and vertical exposures, 2) the lack of high-resolution (sub-million year) stratigraphic correlations between distant outcrop areas, 3) the use of disparate methodologies (i.e. lithostratigraphy, biostratigraphy, event stratigraphy, etc.) and multiple colloquial stratigraphic nomenclatures, and 4) multiple worker and terminological biases both within and between regions. As the Upper Ordovician strata in eastern North America form an important interval in the study of Paleozoic sedimentary geology and an excellent succession for the examination of synorogenic processes, associated depositional signatures, and biotic change across an ancient basin, it is imperative that a series of local, regional and basin-wide correlations be established.

In part, this lack of understanding stems from the need for a regional synthesis and integrated high-resolution chronostratigraphic framework for the analysis of spatially constrained time slices across eastern North America. Moreover, as argued by Dennison (1994), the approach of utilizing tectonic-focused models for explaining potentially complex unconformities and other sedimentary successions without first considering important time-line surfaces and high-order sedimentary fill patterns is a dangerous practice that can dissociate the true chronology of causal patterns.

In this study, detailed high-resolution correlations of regionally-spaced rock successions show the need for a more refined evaluation and empirical testing of the current tectonic models employed for explaining the multi-phased Taconic Orogeny. Through the assembly of a detailed chronostratigraphic model for the cratonic margin foreland basin and the interior epeiric sea, it is possible to establish the lateral extent of erosional patterns associated with unconformities, and patterns and age determinations of basin fill. These data can then be used to investigate the sedimentary response and timing of change within eastern Laurentia. Moreover, given the interplay of eustatic and tectonic processes these data can also be used to establish the chronology and relative impact of both allocyclic and autocyclic events active across the region. Consequently, following the suggestion of Dennison (1994), once eustatic sea-level effects are established and removed from consideration in both foreland and cratonic settings, the residual patterns can then be considered as plausible tectonic signatures and then modeled accordingly. As such, this study has worked to integrate and build on previous stratigraphic studies in order to develop an integrated chronostratigraphy based on biostratigraphic, lithostratigraphic, and sequence stratigraphic analyses–to allow for a more refined analysis of the sedimentary transition out of the GACB into the Taconic Foreland Basin.

OBJECTIVES OF THIS STUDY

As discussed herein, a sequence stratigraphic approach integrating lithostratigraphic, biostratigraphic, and event stratigraphic methodologies has been applied to the Black River-Trenton Group interval (Mohawkian) of eastern North America in order to establish a baseline chronology for evaluating previous observations regarding the dynamics of Taconic Foreland Basin evolution. The Black River-Trenton succession of New York and Ontario and their equivalents elsewhere in eastern North America have become an important testing ground for ideas in basin dynamics of epicontinental seas and foreland basins, sequence stratigraphy, and paleoecology and tectonic evolution (Baird, et al., 1992; Brett & Baird, 2002; Brett et al., 2004; Cornell, 2001; Cisne, et al., 1982; Ettensohn, 1991; Holland & Patzkowsky, 1996; 1997; 1998; Joy, et al., 2000; Lehmann, et al., 1994; Leslie & Bergström, 1995; Mitchell & Bergström, 1994; Rodgers, 1971; Shanmugam & Lash, 1982).

Yet many issues and questions remain unresolved with respect to understanding the scale and scope of the Taconic Orogeny and its impact on the last stand of the GACB. These include:

1) the timing, spatial relationships and scale differences between the southern and northern phases of the Taconic Orogeny;
2) the extent, duration and impact of these distinct tectonic phases on the overall architecture of the Taconic foreland basin, its sub-components and the adjacent GACB;
3) previously unexplained and enigmatic siliciclastic provenance differences between the southern and northern phases of the orogeny;

4) an empirical test for several theoretical basin evolution models, and

5) the relative impact of each of these events on the paleoceanographic and paleoclimatic evolution of the eastern margin of Laurentia and associated implications for biotic change in this region.

GEOLOGIC AND TECTONIC SETTING

The early Upper Ordovician Black River and Trenton groups of northeastern North America consist of between 100 to 500 m of highly fossiliferous carbonates and shales (Keith, 1988). These sedimentary rocks accumulated in subtropical to perhaps temperate climates about 15 to 30 degrees south of the Ordovician equator (**figure 2**). Nonetheless, Mohawkian strata of

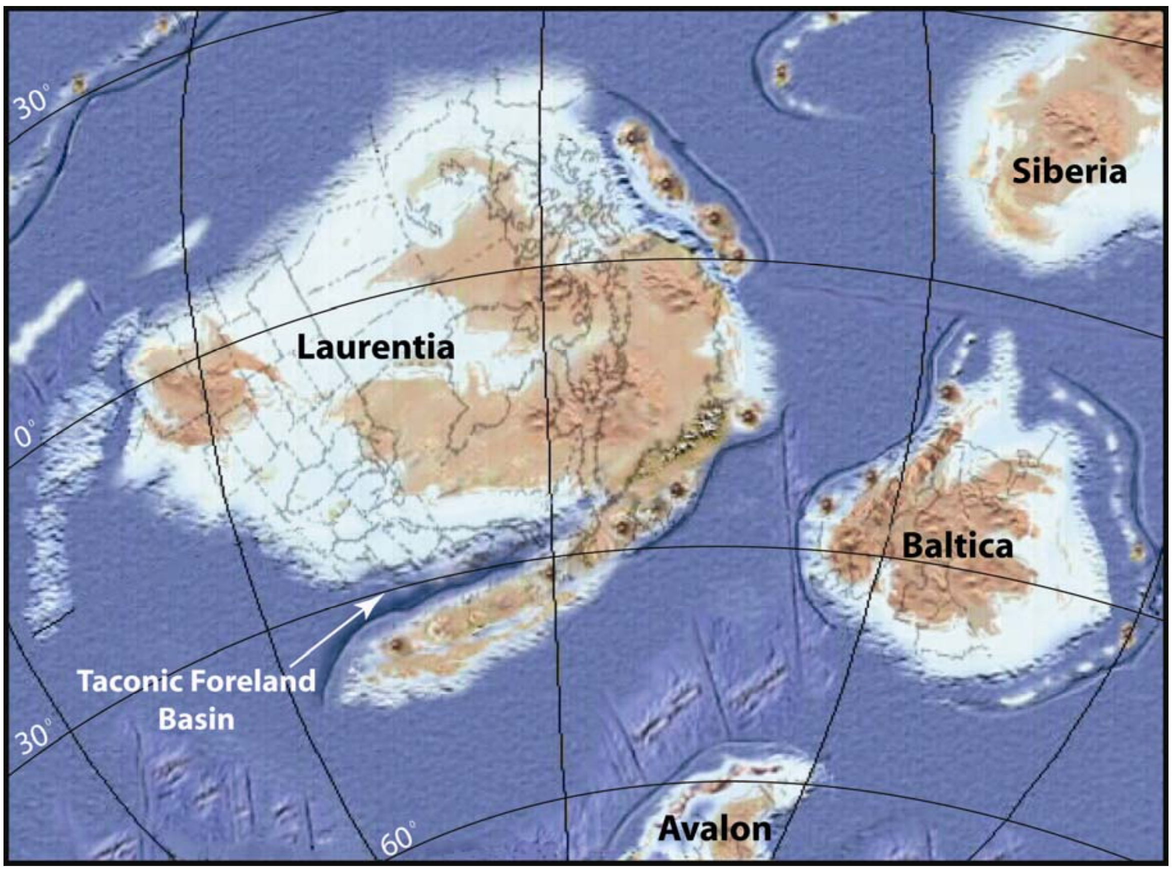


Figure 2: Paleogeographic reconstruction for the peri-Laurentian region during the earliest Late Ordovician. The Great American Carbonate Bank is located to the northwest of the Taconic Foreland Basin as shown above. Figure modified after Blakey (2004).

easternmost North America (earliest Creek Phase of the Tippecanoe Supersequence; Sloss, 1963) record a dramatic and rapid change in depositional setting from widespread carbonate platform environments of the Black River Group into a deep foreland basin during deposition of the upper Trenton Group (see **figure 1**). This transition was heralded by the change from clean, peritidal to shallow shelf Black River carbonates (biomicrites, wackestones, and minor packstones) with coral-dominated faunas, to deep water mixed carbonate-siliciclastic rocks (grainstones, wackestones, and packstones) of the Trenton Group and finally to black shales of the Utica Group (**figure 3**). The sedimentary signatures from these rocks indicate that parts of Laurentia

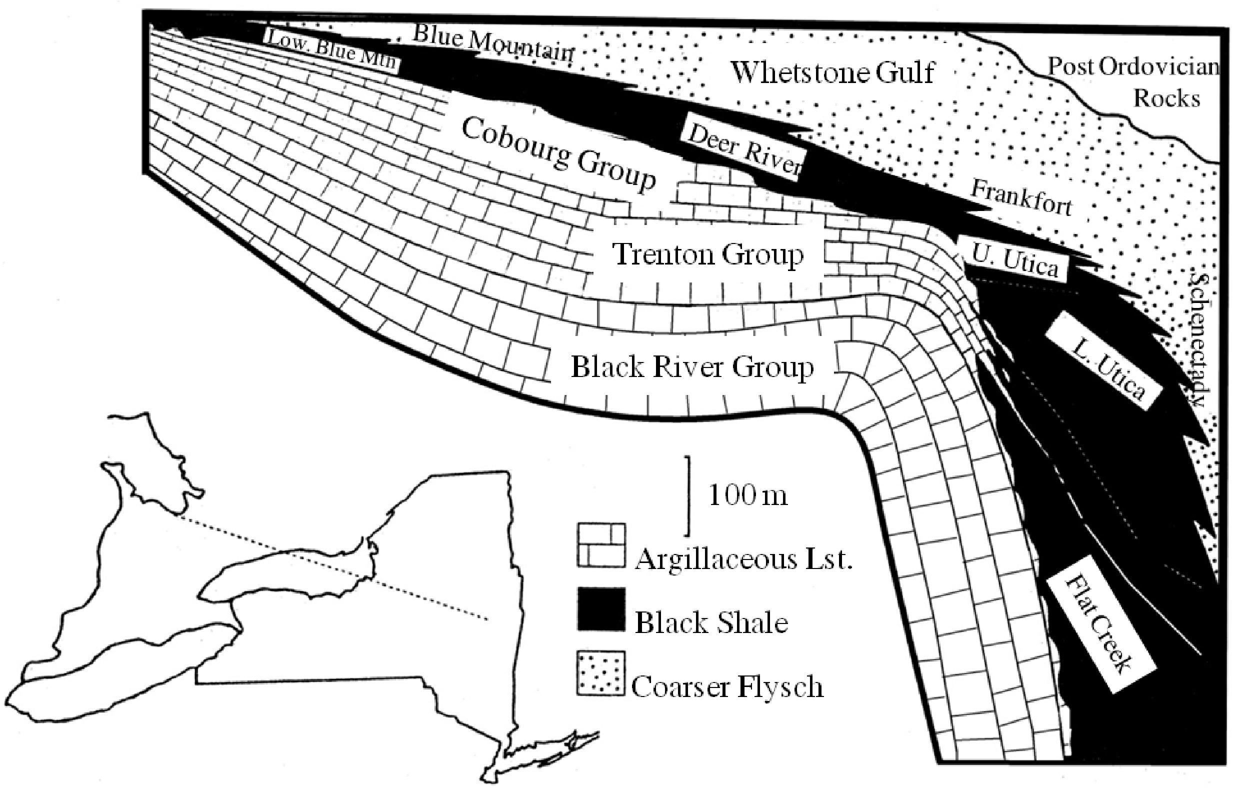


Figure 3: Schematic cross-section for the Ontario-New York Region showing the transition from the Black River and Trenton into the foreland basin deposits of the Taconic Orogeny. Figure modified after Lehman et al., (1994)

not only witnessed substantial topographic alteration through time (**figure 4**), but as a result, the region may also have witnessed substantial oceanographic and climatic alterations in a relatively short interval of time.

Sedimentologic and paleontologic changes recorded in this rock interval have been

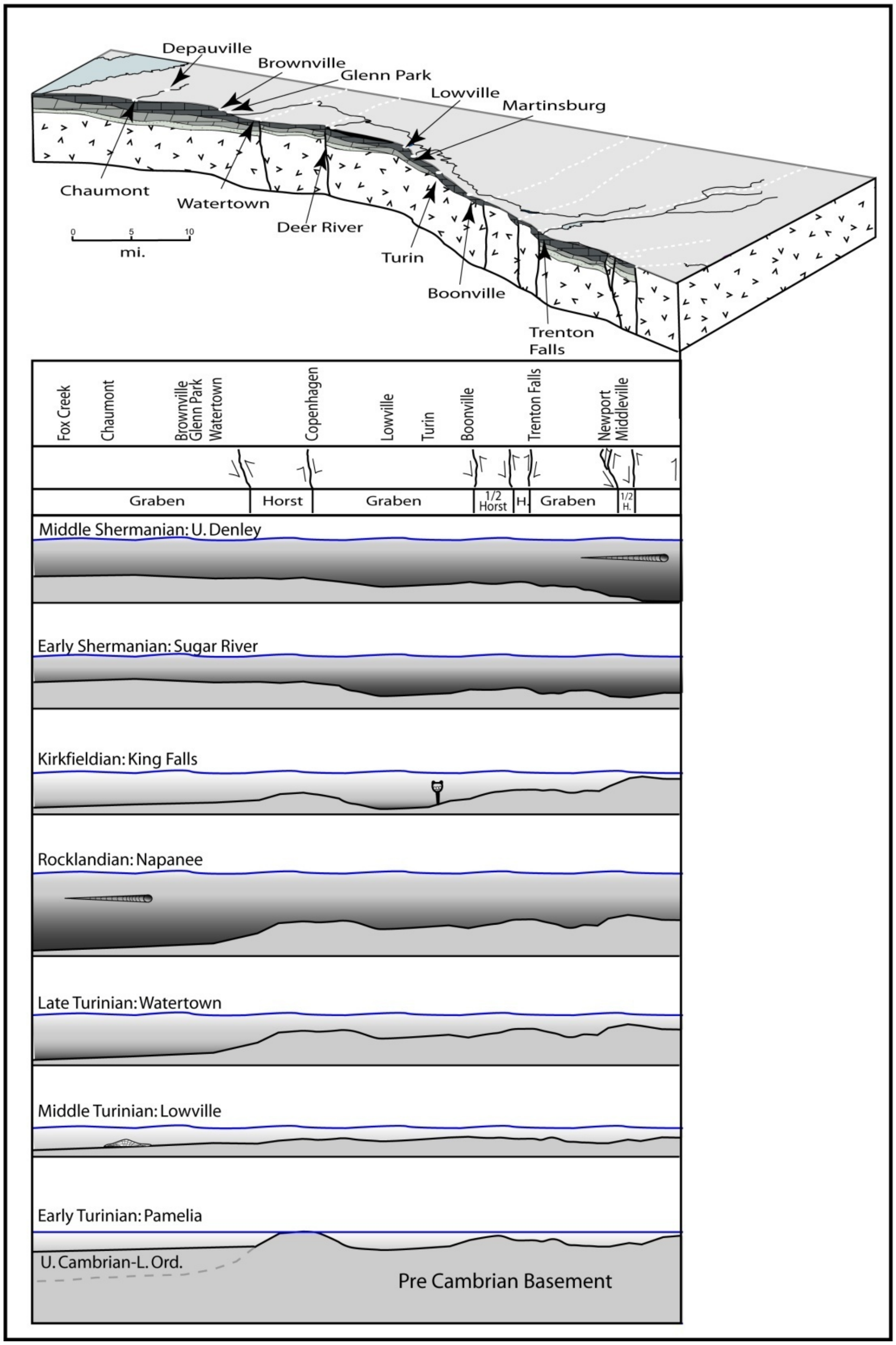


Figure 4: Model for Upper Black River to Lower Trenton Group paleotopographic change through 7 different time series from the Turinian through the Shermanian Stages. Notice the relative reversal of dip direction from that of the protected westward dipping (deepening to the west) carbonate ramp of the Black River to lowermost Trenton, followed by the development of eastward dipping (deepening to the east) ramp to foreland foredeep basin during middle to Late Trenton.

attributed to the collision of an island arc terrain with the eastern edge of North America. The collision is thought to have produced a tectonic load that generated a resultant flexural moat at the Laurentian margin (Fisher, 1962, 1977; Shanmugam & Lash, 1982; Lehmann et al., 1994) and may have also produced more distant flexural highs (forebulges) and lows (back bulge basins) (Holland & Patzkowsky, 1997; Ettensohn, 1994) (**see figure 1**). Overall, tectonic loading is thought to have: 1) resulted in the overall subsidence of the shallow shelf environments pervasive across the GACB during the deposition of the Black River Group, 2) provided a source for the influx of new siliciclastic sediment onto the long subdued craton during and subsequent to the deposition of the Trenton Group, 3) generated both the water depths and basin geometries required to modify water mass circulation patterns given reconstructed paleogeographic and paleotopographic configurations, 4) allowed for the introduction of nutrient-rich water masses that may have impacted sedimentary dynamics during the deposition of the Trenton Group, and 5) influenced the distribution of fossil taxa and their immigration and emigration patterns.

Given the propensity for this tectonic event to have profound influence on the ancestral North American craton and its biota, very little is actually established in terms of high-resolution linkages between the record of tectonic activity, and the record of environmental change during this time. For instance, it is well established from investigations of these rocks that the Taconic Orogeny was not a single event of tectonic change, but a protracted series of events that resulted in the formation of a series of basins and sub-basins originating along the subducted Laurentian margin and within the intracratonic seas (**figure 5**). It has been estimated that Taconic tectonism occurred in at least three tectophases: the first the Blountian Phase activating in the Turinian stage during deposition of the Black River Group. The second, the Vermontian Phase, is thought to have activated early in the Shermanian stage during deposition of the Trenton Group (Cisne et

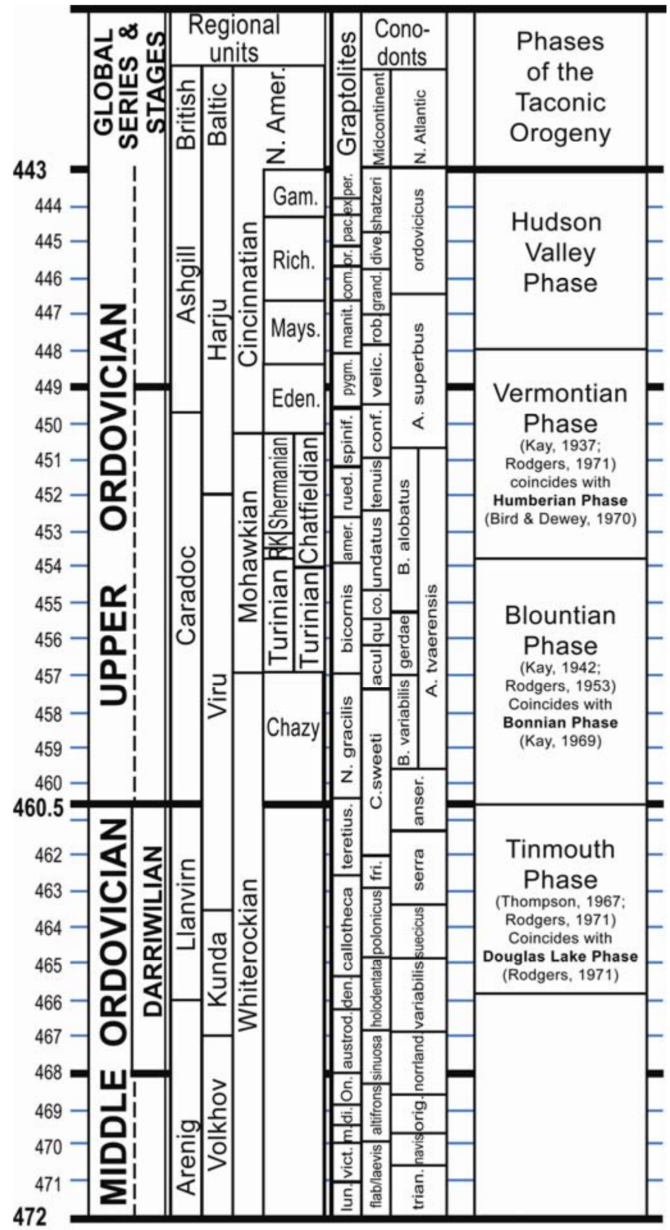


Figure 5: Relative & absolute time scale for the Middle and Upper Ordovician of eastern North America as modified after Webby et. al, (2004). Included are the relative positions of previously identified phases of the Taconic Orogeny including those most commonly recognized as major basin filling episodes which include both the Blountian and Vermontian phases.

al., 1982). The third, reactivated significantly later during the Silurian during deposition of the Medina Sandstone (Ettensohn & Brett, 1998). Each tectophase resulted in a number of modifications to the structural and topographic expression of eastern Laurentia.

Regardless, it is quite clear that the pattern of sedimentologic and stratigraphic change in the epicratonic seas was impacted by local and regional tectonism as well as global eustatic sea-level change. What is not yet clear, however, is the degree to which these structural modifications impacted sedimentation patterns and therefore a number of environmental constraints including water mass circulation patterns, climatic regimes, and biogeography in the region.

GEOLOGIC TIME SCALE

The Chazy, Black River and Trenton groups of New York State are among the earliest formally designated stratigraphic units in North America (Emmons, 1841; Vanuxem, 1842). The names of these well-known intervals have been applied throughout much of eastern North America (for examples see Ulrich, 1911; Butts, 1940; Kay, 1960). These units were originally described by a combination of their lithologic and paleontologic composition. Subsequent investigations, therefore, considered these not just as distinctive rock-units, but also as distinctive biostratigraphic units. Therefore they were equated with time-rock units and used to denote specific intervals of time (i.e. Chazyan, Black Riveran, and Trentonian) (**figure 6**). The practice of using the same nomenclature for rock-terms and time-rock terms has been challenged on the basis that rock units themselves should not be interpreted as indicative of any specific time. In some cases rock units (specific lithologies) may be diachronous through time according to Walther's Law and therefore they should not be used to convey any formal time position themselves. Not surprisingly, stratigraphers have discouraged the use of lithostratigraphic terms for time-rock terms. Therefore the older rock terminology (i.e. Chazy, Black River, Trenton), although in common use today as time-rock units (Chazyan, Black Riveran, and Trentonian), are strictly to be used as descriptors of lithology. New time-rock terms have thus been constructed

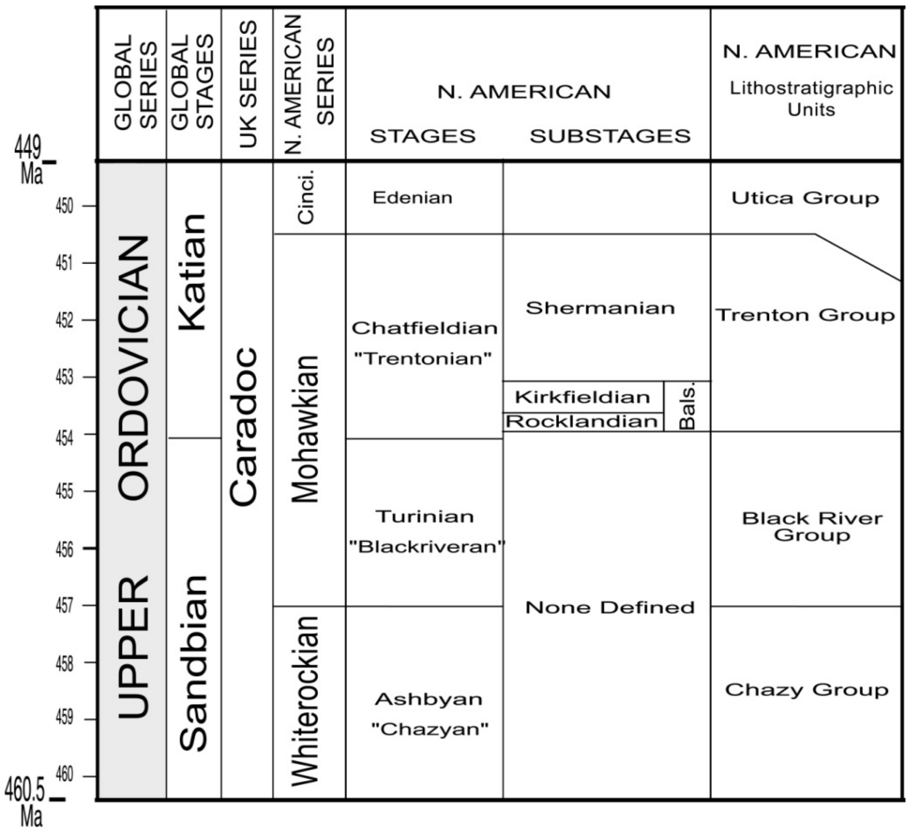


Figure 6: Geologic time scale for the Upper Ordovician Series of eastern North America. Included are the newly recognized Global Stages (Sandbian and Katian). Also shown are the positions of North American stages and substages of previous authors as well as the lithostratigraphic positions of key Upper Ordovician units of the study regions. Abbreviations: Cinci. = Cincinnatian; Bals. = Balsamian.

to specify time-stratigraphic positions for chronologic study.

Among the first attempts at separating time-rock from time-independent rock terminologies were the stage level classifications of Kay (1937, 1960). Among these stage level classifications were the Rocklandian, Kirkfieldian, and Shermanian stages. During these intervals collectively, the Trenton Group rocks were deposited. More recently, however, due to difficulty in recognizing Kay's three upper Mohawkian stages outside of New York and Ontario, the interval that was called the "Trentonian" elsewhere has been replaced by the Chatfieldian Stage for Chatfield, Minnesota with the base established at the position of the Millbrig K-bentonite (Leslie & Bergström, 1995). Leslie and Bergström's term has now been adapted by many workers and has been included in most recent Ordovician time-scale calibrations (Cooper

& Sadler, 2004). Kay's terms are thus used as sub-stages of the Chatfieldian, with the exception that Barta and colleagues (in press) have suggested the replacement of the Rocklandian and Kirkfieldian with the term Balsamian for the region around Balsam Lake in southern Ontario – close to the type Kirkfieldian.

Kay (1948) attempted to provide a time-rock equivalent nomenclature for the "Black Riveran" and therefore established the Bolarian series, with the Hatterian, and Hunterian sub-stages. These were named from units he described in central Pennsylvania based on their lithologic (and associated biostratigraphic signatures) similarity to New York sections. However, these stratigraphic terms, erected from less well studied outcrop sections, were not readily incorporated into academic discourse without independent biostratigraphic support for time correlation with the type region even though Kay continued to use them (Kay, 1960; 1968).

To add to the dilemma Cooper (1956), in his mega synthesis on "Chazyan and Related Brachiopods," added fuel to the already difficult time-rock battle. Cooper's approach, rather than using collective lithostratigraphic and biostratigraphic evidence, was to institute a time-rock classification for this interval on the basis of brachiopod occurrences. Recognizing distinctive assemblages of brachiopods as either Trenton-like, Black River-like, or Chazy-like, Cooper established the terms Porterfield[ian] (for basal Black River units), and Wilderness[ian] (for upper Black River to lower Trenton units). Although receiving more support on their biostratigraphic basis, Cooper's (1956) terms did not equate well with the long-standing status of New York type sections and were therefore rarely used. Due to these complexities, Kay and Cooper's terminologies were abandoned and Fisher established the term Turinian for outcrop sections in the central Black River Valley of New York State. Again, the addition of this new nomenclature presents difficulty with correlation outside of New York, but roughly corresponds to the entire *Climacograptus bicornis* graptolite biozone with the base of the Turinian established at the base of the *Belodina gerdae* conodont sub-zone (Kolata et al., 1996) and so it has generally been accepted (Cooper & Sadler, 2004).

Today, the Turinian and Chatfieldian Stages make up the Mohawkian Series of the North American time-rock classification (Cooper & Sadler, 2004). Nonetheless, the alternative stage classification, using Kay's (1937) more precise subdivisions of the upper Mohawkian Series: the Rocklandian, Kirkfieldian, and Shermanian stages, are still useful in the type region and perhaps more useful than the term Chatfieldian. Moreover, these terms are also still in common use in neighboring regions, and are now recognizable in sections further afield, by correlation independent of biostratigraphy. Therefore their usage is followed herein in order to facilitate discourse with older literature and provide a basis for dialogue with traditional stratigraphic practice established by the Stratigraphic Code.

The Chazy Group underlies the Trenton-Black River Groups. In similar fashion as the term "Black Riveran," the Chazy Group rocks of the Lake Champlain region of New York and Vermont have also been co-utilized in common parlance to refer to both the rock units themselves and the time of deposition. Although Kay (1948) commonly used Ulrich, Raymond and Grabau's concept of the term "Chazyan" (Shaw, 1969), he suggested that suspected time equivalent rocks, to the Chazy Group from southwestern Virginia, could supply proper time-rock terms. He therefore suggested informally the terms Blackford-Five Oaks (for lower Chazy) and Lincolnshirian (for upper Chazy) pending further investigation. Building on Kay's informal terms, Cooper (1956) recognized a brachiopod-based biostratigraphic zonation within the "Chazyan" and suggested a formalized stage nomenclature and suggested using Marmor and Ashby[an]. Realizing the inadequacy of his earlier classification, Kay (1960) returned to using type-region nomenclature for the time-rock terminologies. In his “Classification of the Ordovician System of North America” volume, Kay attempted to elevate the Chazyan to series stature, and proposed the Dayan, Crownian, and Valcourian as stages of that series. Without substantial new biostratigraphic support at the time for Kay’s “Chazyan Series,” Fisher (1962) followed Cooper’s (1956) classification for the type-section and established the correlation of Kay’s upper Day Point and Crown Point (Dayan and Crownian) with Cooper’s Marmor, and Kay’s Valcour (Valcourian) with Coopers’ Ashby.

Although some argument still remains as to the exact correlation of Chazy type section rocks with equivalents in the central and southern Appalachians and elsewhere (Shaw, 1968; 1969), recent biostratigraphic analysis suggest all units of the type Chazy are younger than Cooper’s Marmor and therefore the type “Chazyan” is mostly Ashbyan in age (Sweet & Bergstrom, 1976; Finney, 1982; Sweet, 1984). The Chazy of New York is thus coincident with *N. gracilis* graptolite zone and the upper *C. sweeti* to lower *P. aculeata* and lower *A. tvaerensis* conodont zones (Webby et al., 2004). Sweet’s (1984) conodont biostratigraphy has thus helped to establish the position of the Chazy as very latest Whiterockian Series, Ashbyan Stage. Thus the contact of the Chazy with the overlying Black River Group is coincident with the Whiterockian-Mohawkian Series boundary.

It should be noted that the IUGS Subcommission on Ordovician Stratigraphy has recommended that the interval previously termed Middle and Upper Ordovician (i.e. the Llandeilo, Caradoc, & Ashgill stages) all be assigned to Upper Ordovician (Webby, 2004). Thus, the interval of time during which the Chazy, Black River and Trenton groups were deposited now lies entirely within the Late Ordovician, and NOT the Middle Ordovician as previously defined. In the British classification the Upper Ordovician, Whiterockian-

Mohawkian interval is thus assigned to the Caradoc Series, and is coincident with Webby and colleagues' 5a through 5d time slices.

Absolute age dating for this interval has become a more realistic endeavor in recent years as multiple K-bentonites have yielded datable phenocrysts both inside and outside of the New York region (Samson et al. 1988, Haynes, 1994; Kolata et al., 1996, Min et al., 2001; McLaughlin et al., 2005; McLaughlin et al., in press). Two widely distributed K-bentonite horizons, the Deicke and Millbrig, have been dated by a number of geochronometric systems. The resulting ages using U/Pb dating methodology are established at about 454.5 +/- 0.25 Ma for the Deicke K-bentonite; and to 453.1 +/- 0.65 Ma (for U/Pb dates of zircon) for the Millbrig (Min et al., 2001). Moreover, recent dating of the High Falls K-bentonite (from Trenton Falls) has yielded a highly concordant U/Pb zircon date of 451.8 +/- 0.3 (McLaughlin et al., 2004; McLaughlin et al. in prep). Based on these assessments and by extrapolation from dates on the Ordovician-Silurian boundary, the entire Black River to Trenton Group interval ranges from approximately 457 to 450 million years providing for 6-7 million years of deposition. Inclusive of the Chazy, the Ashbyan Stage, the entire Mohawkian Series, and the lowermost Cincinnatian, this interval spans roughly 10 million years of deposition.

GEOGRAPHIC SETTING OF ORDOVICIAN OUTCROP EXPOSURES

Ordovician outcrop exposures are widespread throughout eastern North America (**figure 7**). Following the precedent of Kolata and colleagues (1996), there are approximately 15 localized areas of outcrop exposure in the eastern half of the continent ranging in a 1500 km wide swath from western Texas northeastward to Newfoundland, across a distance of nearly 4000 km. Although fairly substantial in extent, the portion of these outcrop regions exposing

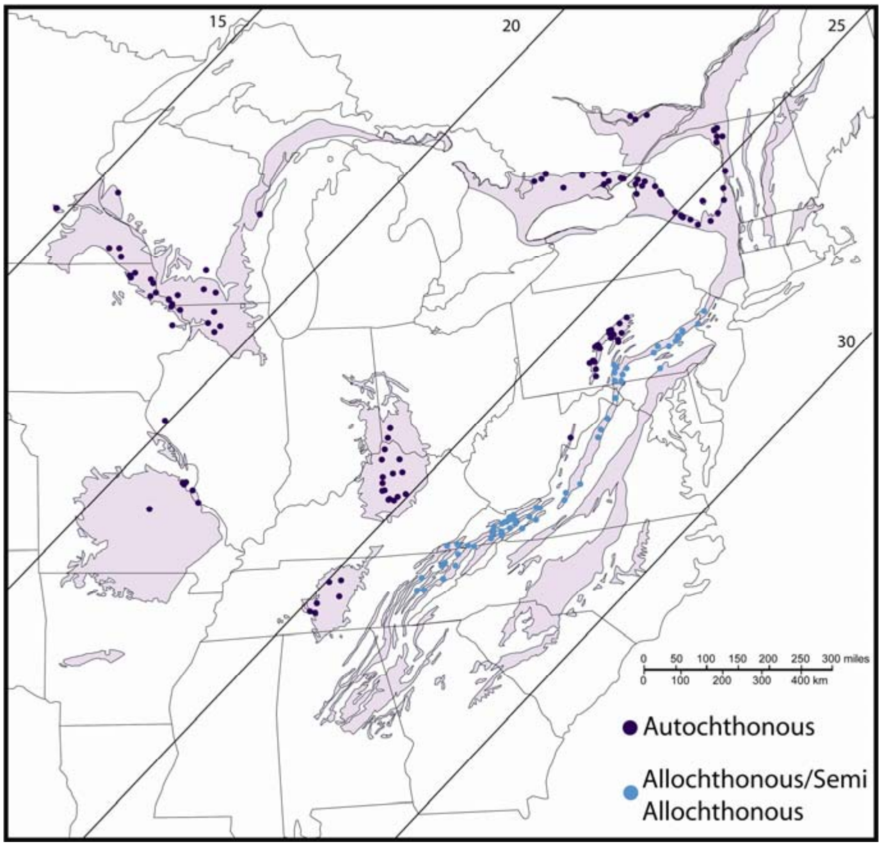


Figure 7: Ordovician outcrop distribution in eastern North America (shaded) with location of type localities for Mohawkian strata (dots), distinguished by their structural context (i.e. allochthonous – indicates outcrop localities that have been disturbed tectonically and have been folded, faulted and transported to some extent from their original depositional location). Relative paleolatitudinal position (in° South) is also demarcated after Scotese and McKerrow (1990), and Witzke (1990).

Upper Ordovician rocks is substantially less with most Upper Ordovician exposures limited to the periphery of larger structural regions.

The largest, well-exposed, succession of Upper Ordovician rocks occurs only within the mid-continent region along the Cincinnati Arch/Jessamine Dome, and to a lesser extent in the Nashville Dome, and the Adirondack Lowlands/Grenville Shield region. Additional important exposures are also known from intracratonic settings in the U.S. from the Upper Mississippi River Valley along the Wisconsin Arch in Wisconsin and neighboring southeastern Minnesota and northwestern Iowa. Other exposures located in the marginal cratonic belts include those from the southern Appalachians in eastern Tennessee/southwestern Virginia, the central Appalachians in northern Virginia/western Maryland/south-central Pennsylvania, and in the northern

“Appalachians,” in northeastern New York/Vermont; and finally in the St. Lawrence River Valley northeast of the Adirondack Lowlands. Additional exposures are also known from the marginal cratonic Maritime Provinces of Canada including Nova Scotia and Newfoundland.

For the purposes of this study, the main outcrop regions considered in most detail are focused on the intracratonic regions of the Cincinnati Arch, the Ontario-Northern New York region, and the Valley and Ridge province (near-marginal to marginal-cratonic successions) in south-central Pennsylvania, and eastern New York State. For continuity with other well-studied outcrop sections, some attention is also paid to the Upper Mississippi Valley and the Nashville Dome (both intracratonic settings), as well as the eastern Tennessee and Virginia Valley and Ridge sections (marginal cratonic setting).

PALEOGEOGRAPHIC SETTING

Lithologic and paleobiologic indicators suggest that the Chazy, Black River, and Trenton Group sediments of eastern North America accumulated in subtropical to sub-temperate latitudes during the Upper Ordovician. Paleomagnetic studies indicate that Laurentia was located in a position straddling the paleoequator during this time (Scotese & McKerrow, 1990; McKerrow et al., 1991). Cratonic sequences of the GACB of eastern Laurentia were therefore located in the southern hemisphere such that most of the areas exposed in the modern outcrop belt were located, during the Ordovician, in subtropical latitudes ranging to about 30°S of the paleoequator.

Latest Whiterockian and mid-Mohawkian strata from marginal cratonic settings are dominated by shallow-water tropical carbonates throughout much of the Ordovician. In the south and north respectively, these are documented to record a tectonically induced change from widespread shallow carbonate platform environments of the CAGB into actively subsiding, shale-dominated basins. Moreover, due to the onset of Sloss's (1963) Tippecanoe megasequence **(figure 8)** in the Late Upper Ordovician, sea-levels were also, over the long-term, in a

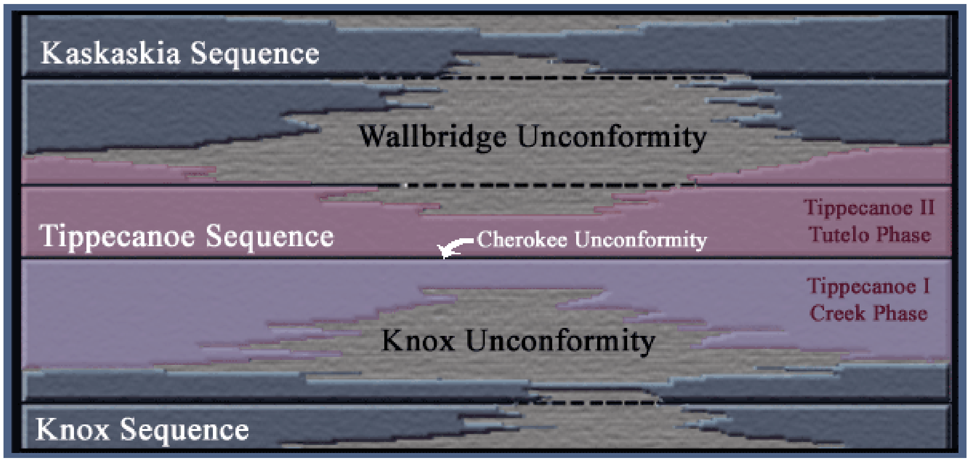


Figure 8: Sloss's Megasequences (1963) for the Lower Paleozoic and associated unconformities. The Tippecanoe sequence is shown labeled with its early and later phases which are shown separated by the Cherokee Unconformity formed during the end Ordovician Hirnantian Glaciation when sea-levels were significantly reduced from previous highstand positions. Figure from Cornell et al., 2004.

transgressive phase such that during the Cincinnatian, very little land was exposed on the continent. In fact, the only exposed areas on the continent were portions of the Canadian Shield and areas along the Transcontinental Arch which were of very low relief (**see Figure 2**). Although there are no modern epicontinental seas comparable in size to the GACB in the Cambro-Ordovician, modern regions such as Florida Bay, the Bahamian Platform, and the Persian Gulf are commonly used as analogs for understanding the depositional conditions of the Laurentian craton at that time. Moreover, Lash (1984, 1986) proposed a modern analog for the tectonic setting of the eastern cratonic margin of Laurentia during the Ordovician. The modern northwest Australian shelf is known to be overthrusted by slices of the Timor accretionary wedge (Kaneko, et al., 2007). In this scenario, a modern elongate basin has developed to the southeast of the Timor – Tanimbar island chain. This elongate basin, known as the Timor trough, has a NE-SW axial orientation **(figure 9)**. The oblique collision has induced the depression of the

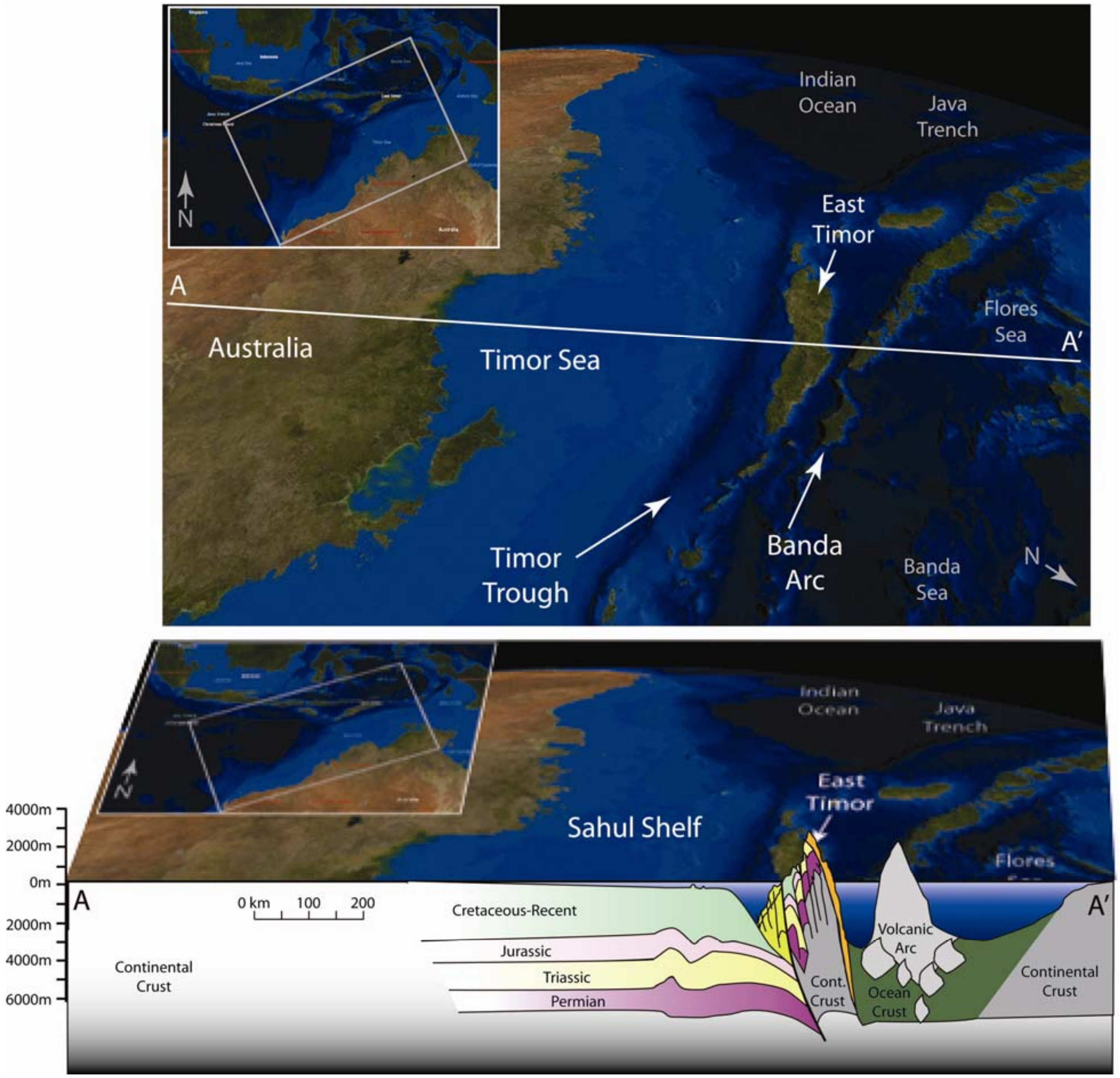


Figure 9: Modern geographic map and schematic tectonic model for the north-west Australian shelf collision with the Eurasian Plate. The positions of the Timor Trough, the Sahul Shelf, and the Timor Accretionary Arc are shown as well as the Banda Volcanic Arc. View for the lower schematic cross-section (transect A-A') is to the southwest towards the Indian Ocean along the axis of the Timor Trough. Map image capture and modified from NASA World Wind software, and schematic cross-section is adapted from Norvick (1997).

trough, by the tectonic load of the associated accretionary wedge (Timor). In addition to the exposed non-volcanic accretionary island chain, there is a volcanic island arc located to the north and is referred to as the Banda Arc. It appears from recent investigation that although the Australia Plate is moving to the northwest at a rate of between 6-8 cm/year, subduction activity along this region has ceased and tectonic activity is currently limited to isostatic doming in the Timor region as well as reverse faulting along the Wetar Thrust that may be a signature of the initiation of subduction reversal (Kaneko, et al., 2007).

PLATE TECTONIC SETTING AND STRUCTURAL FRAMEWORK OF LAURENTIA

Classically studied for over a century, the geologic condition of the eastern margin of North America has long been established as resulting from a number of re-occurring yet inter-related pulses of sedimentation, deformation, and metamorphism (Dewey, 1969 a,b; Bird & Dewey, 1970). Collectively termed the Appalachian Orogen, Bird and Dewey (1970) recognized evidence for an extended period (Ordovician through Devonian [and ultimately through Pennsylvanian]) of contraction in the proto-Atlantic region. During this time repetitive periods of plate accretion and continent margin subduction ultimately modified the passive margin of eastern Laurentia. Their seminal study helped to introduce the concept of plate tectonics into the long-standing “Taconic problem” through the suggestion that the changes in sedimentation, patterns of deformation, metamorphism, and even volcanism observed in the Appalachians could be related to the initiation of subduction along the eastern cratonic margin. This was considered to be the case especially during the Taconic Orogeny in New England and the Humberian Orogeny of Newfoundland (Bird & Dewey, 1970). Although their suggestion of a west-dipping subduction zone is generally not excepted today (but see Karabinos et al., 1998), these authors along with Rodgers (1971) helped to reinvigorate the outcrop-based investigation of the “Taconic problem” and a flurry of new theoretical studies of continental margin deformation and lithospheric flexure was initiated (see, for example: Church & Stevens, 1971; Chapple, 1973;

Rowley et al., 1979; Beaumont, 1981; Rowley & Kidd, 1981; Cisne et al., 1982; Shanmugam & Lash, 1982; Bosworth & Rowley, 1984; Quinlan & Beaumont, 1984).

Following the studies of Church and Stevens (1971), and Chapple (1973), who presented data that indicated an east-dipping subduction zone (in contrast to the west-dipping model of Bird & Dewey, 1971) for the Newfoundland Humberian Orogeny, Rowley and colleagues (1979) and Rowley and Kidd (1981) presented a stratigraphic argument which suggested the Taconic allochthons were detached from their basement through obduction and accretion along an east-dipping subduction zone. Although their model is generally accepted today, Karabinos and colleagues (1998) have recently suggested that the subduction direction may have indeed been both east and west. That is east-dipping subduction in the Rowley and Kidd model was followed by subduction reversal to west-dipping (Bird and Dewey model) after Laurentian continental crust failed to subduct under the proto-Atlantic (Iapetan) oceanic crust.

Data presented by Karabinos and colleagues (1998, 2001) designate that the Shelburne Falls Volcanic Island Arc (absolute age dates of 485-470 mya) collided with the eastern margin of Laurentia along an east-dipping subduction zone – a condition similar to that of present day NW Australia. This was followed by subducted slab break-off, subsequent reversal to west-dipping subduction, and the development of a second volcanic arc: the Bronson Hill Arc (454-442 mya) located outboard of the previously accreted Shelburne Falls Arc.

Although these data will be further considered in a subsequent discussion, the Karabinos model has some opponents including Ratcliffe and colleagues (1999). Nonetheless, in order to accommodate complete closure of the Iapetus Ocean, during the Devonian collision of the Avalon micro-continent in the Acadian Orogeny, west-dipping subduction had to have been initiated along this margin at some point. If subduction did not reverse from the east-dipping scenario to west-dipping during the Taconic Orogeny, it is not clear when this event would have happened. Moreover, there are data that suggest that after just under 5 million years of collision, the ongoing modern Timor-Australian orogeny is indeed undergoing slab break-off leading to subduction zone reorganization (Richardson & Blundell, 1996; Kaneko, et al, 2007).

Despite these arguments, by the time the Chazy and Black River were deposited, and certainly by the time the Trenton limestones were deposited, the eastern margin of Laurentia had undergone major paleotopographic changes. These changes quite obviously led to the reorganization of paleoenvironments as well as paleoceanographic circulation patterns in the peri-Laurentian region. Evidenced by the shift from widespread, clean, peritidal and sub-tidal carbonates of the Chazy and Black River groups, to shale-prone, and even shale-dominated, deeper water, off-shore deposits of the Trenton Limestones and Indian Castle shales respectively, these changes point to significant tectonic modification of the eastern Laurentian margin. The transition from a passive drifting continental margin to an active collisional one is associated with the uplift and subsidence of portions of the Laurentian craton beginning in the Ashbyan. Thus initiated, Taconic tectonism ended a period of over 30 million years (Late Cambrian to Mid. Ordovician) where relatively little tectonic activity occurred on the GACB (Read, 1989). However, in the 10 million year period that ensued the Laurentian margin (along the present day east coast of the United States) was impacted by three episodes or pulses of lithospheric modification (in the Late Ordovician and perhaps more in the Silurian) associated with active collision within the Taconic Orogeny (**see figure 5**; Rodgers, 1971; Ettensohn, 1994; Ettensohn & Brett, 1998; Ettensohn, et al., 2002).

The first obvious tectophase, the Blountian, is defined on the basis of rapid subsidence of continental margin areas showing evidence of sediment starvation, mineral enrichment, and black shale deposition. This event is located far south of the type area of the Taconic Orogeny and the Taconic Mountains of eastern New York. This southern and earlier phase occurred in the vicinity of eastern Tennessee and impacted sedimentation in the Sevier Basin in western North Carolina and eastern Tennessee and neighboring regions beginning in the Ashbyan with flysch deposits and continuing through the Turinian Stage when this basin became overfilled with molasse **(figure 10)**. In the autochthonous successions of the Taconic region, the impact of

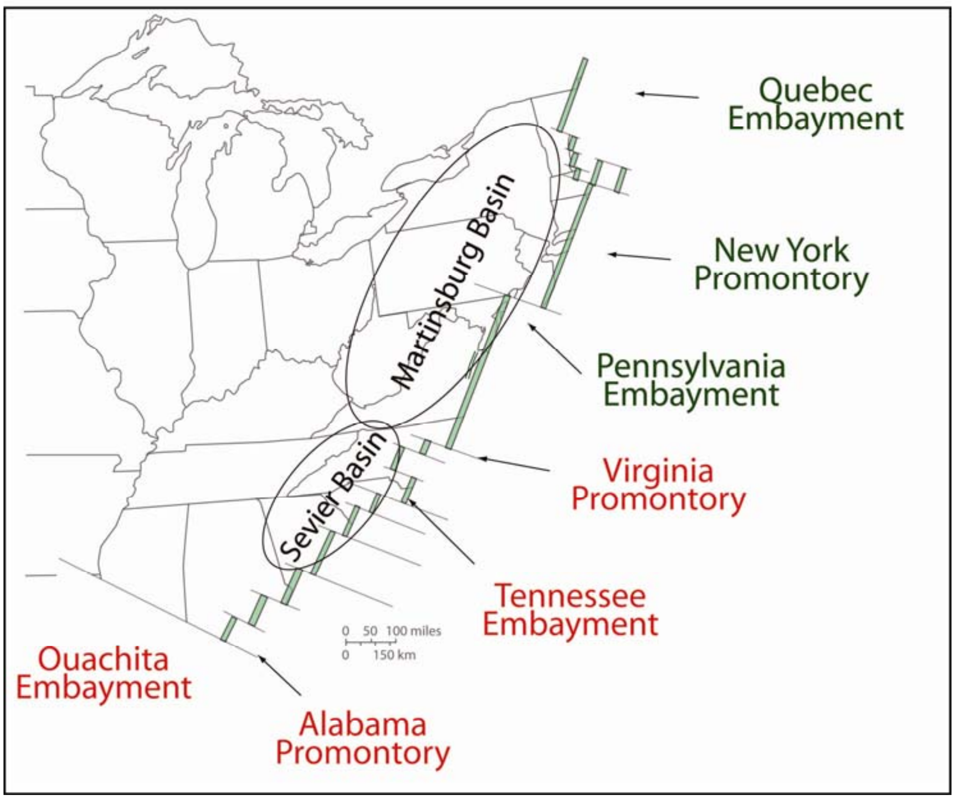


Figure 10: Paleogeographic map of eastern North America showing the position of the Ordovician cratonic margin as recognized from deep seismic studies (modified from Thomas, 1991). Also shown are the positions of key promontories and embayments along the ancestral margin. Shown in red are the locations of promontories and embayments that were likely impacted by initial collision during the first or Blountian Phase of the Taconic Orogeny forming the Sevier Basin. Likewise shown in green are the features involved in the second or Vermontian phase of the Taconic Orogeny.

Blountian-aged siliciclastic sedimentation is generally lacking on the craton. However a dark grey shale body is recognized in the shelf successions in the Crown Point Formation from Plattsburgh, NY and Isle LaMotte, VT in the time equivalent Chazy Group. This autochthonous shale unit is most likely a cratonward extension of the allochthonous Upper Mt. Merino black shales which were deposited in continental slope and oceanic floor settings. These deep water deposits are constrained biostratigraphically to the *N. gracilis* zone (of Berry, 1960; Riva, 1968, 1974; Rickard & Fisher, 1973). These are also coincident with the time equivalent Blockhouse black shales of the Sevier Basin (Finney et al., 1996). Whether tectonic or eustatic in origin, the coincidence provides, at the least, a correlation between the southern and northern cratonic margin successions at this time.

The second and latter tectophase, referred to as the Vermontian Tectophase, was a much more substantial event. This event impacted the vicinity of the New York Promontory producing the Taconic Foreland Basin or Champlain Trough of Kay (1960), and adjacent intracratonic Martinsburg Basin. Although its initiation is the subject of much debate, and a focus of this study, it is clear that the initiation of the Taconic Orogeny was well underway during the deposition of the Trenton Group in the Upper Mohawkian. Ultimately, this later event impacted a much broader region of eastern North America and influenced sedimentary changes across the entire eastern Laurentian craton. These sedimentary changes are recognized in the Michigan Basin (Coakley & Gurnis, 1995), and as far west as Minnesota, Arkansas and Oklahoma as supported by Nd Isotope studies (Fanton et al., 2002; Gleason et al., 2002). Moreover, as discussed herein, due to the much larger scale of this latter tectophase, the Vermontian event significantly shrank the areal extent of carbonate production and effectively led to the demise of the Upper Ordovician GACB.

Although the relationships of these tectophases have been studied for some time (Bird & Dewey, 1970; Rodgers, 1971; Ettensohn, 1991, 1994), many circumstances of their spatial-temporal relationships and dynamic impact on the Laurentian craton are still being worked out, especially in light of provenance studies (Mack, 1985; Andersen, 1995; Bock et al., 1998).

Basinal siliciclastic sediments were dominated initially by muds, but in later phases of each tectophase these fine-grained units were replaced upward and laterally by silts (siltstones), sands (sandstones), and even gravels (conglomerates) as the rate of tectonic loading decreased and the foreland basins were filled in.

Although both tectophases produced similar basin fill packages (Shanmugam & Lash, 1982), provenance studies of these siliciclastic sediments have highlighted a major difference in the source of Blountian and Vermontian sediments (Mack, 1985; Andersen, 1995; Bock et al., 1998). These data from both sandstone petrography and mudrock geochemistry, suggest Blountian sediments were eroded from sources that were derived from the Laurentian cratonic margin. It is likely that these represented portions of the accretionary prism that were uplifted during the collision of the Blountian tectophase (**see figure 5**).

On the other hand, siliciclastic sediments deposited during the Vermontian tectophase (including those of the Trenton Group) indicate a significant shift in provenance. Sediments both on the craton and in the foreland basin show mafic signatures that were likely derived from volcanic source rocks. This dichotomy indicates a fairly major, yet unexplained, difference in tectophase dynamics between the earlier and later tectophases. On this basis it is important to reevaluate all previous models for the Taconic Orogeny in the eastern United States in order to consider these important data. It is reasonable to accept that the development of orogenic activity in eastern Laurentia resulted from the development of subduction along the Iapetan Oceanic Plate. However the exact timing and mechanism for collision has yet to be explained. Nonetheless, this tectonic activity likely involved the following:

Blountian Phase:

- Tectonic loading of the margin from the piggyback development of older slope and rise and ocean floor deposits in a thrust complex (likely an accretionary wedge),
- Cratonic margin subsidence under the newly emplaced load,
- the development of a peripheral foreland basin or deep-water trough along the margin of the once shallow cratonic margin,
- the delivery of basin-filling siliciclastic sediments derived from recently uplifted source terranes in the collisional belt into a narrow foreland basin of limited lateral extent.

Vermontian Phase:

- subsequent and rejuvenated episodes of tectonic shortening shifted subsidence further to the northeast with westward migration of the foreland basin and associated features (foredeep, peripheral bulge) through time,
- new volcanic sediment sources began to contribute both fine and coarse grained sediments into the foreland basin as well as out into the epeiric sea.
- emplacement of allochthonous materials on top of older passive margin deposits and early syntectonic deposits.

Additional evidence for tectonic activity comes from altered volcanic ash layers deposited on the carbonate platform as well as within the foreland basins. Beginning early in Black River deposition and continuing throughout the Mohawkian, there are numerous K-bentonites (altered volcanic ash layers) that provide direct evidence for explosive tectonic activity in a nearby region. The predominance of these K-bentonites supports the theory of a volcanic island arc located offshore from the southeastern margin of Laurentia at this time. These volcanic terranes were responsible for some of the largest volcanic eruptions known in earth's history (Kolata et al., 1996; Huff et al., 1996). These volcanic terranes, i.e. the Ammonoosuc Arc of Rowley & Kidd (1981), the equivalent of the Bronson Hill and Shelburne Falls Arcs of Karabinos and colleagues (1998), the Chopawamsic-James Run Volcanic Arc (Evans, 1984), etc. **(figure 11)**,

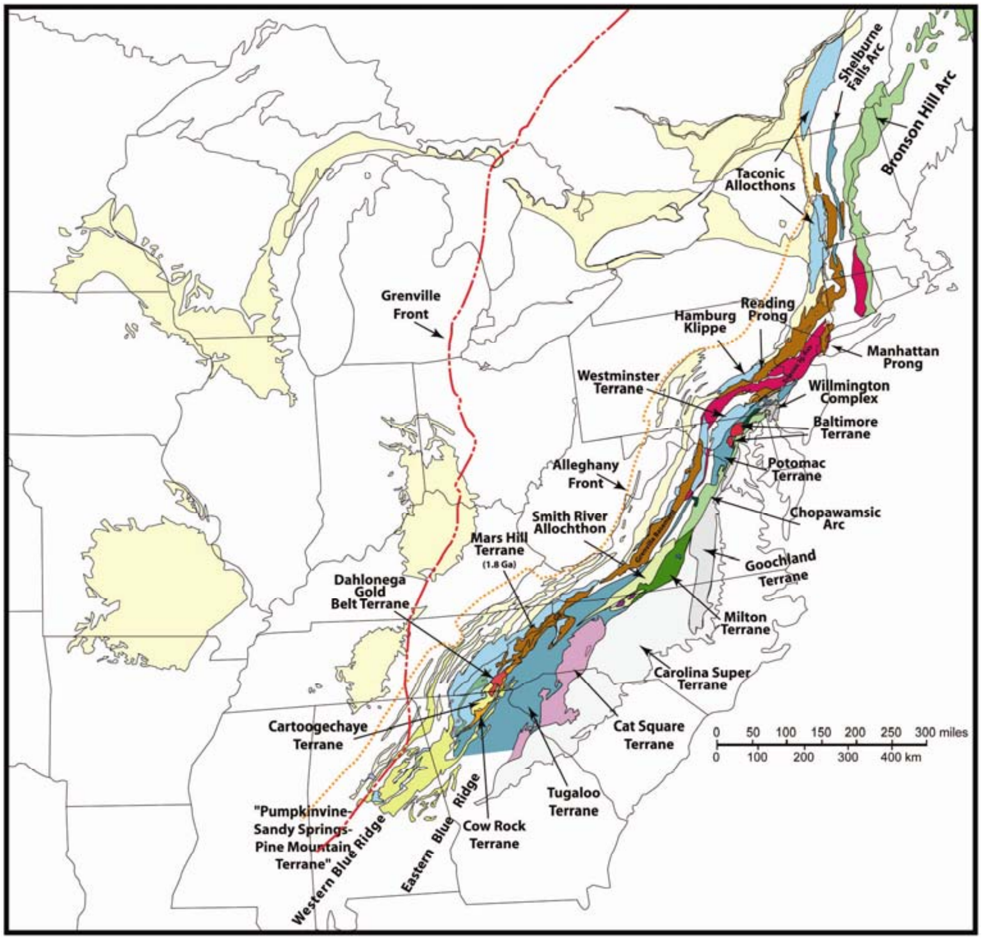


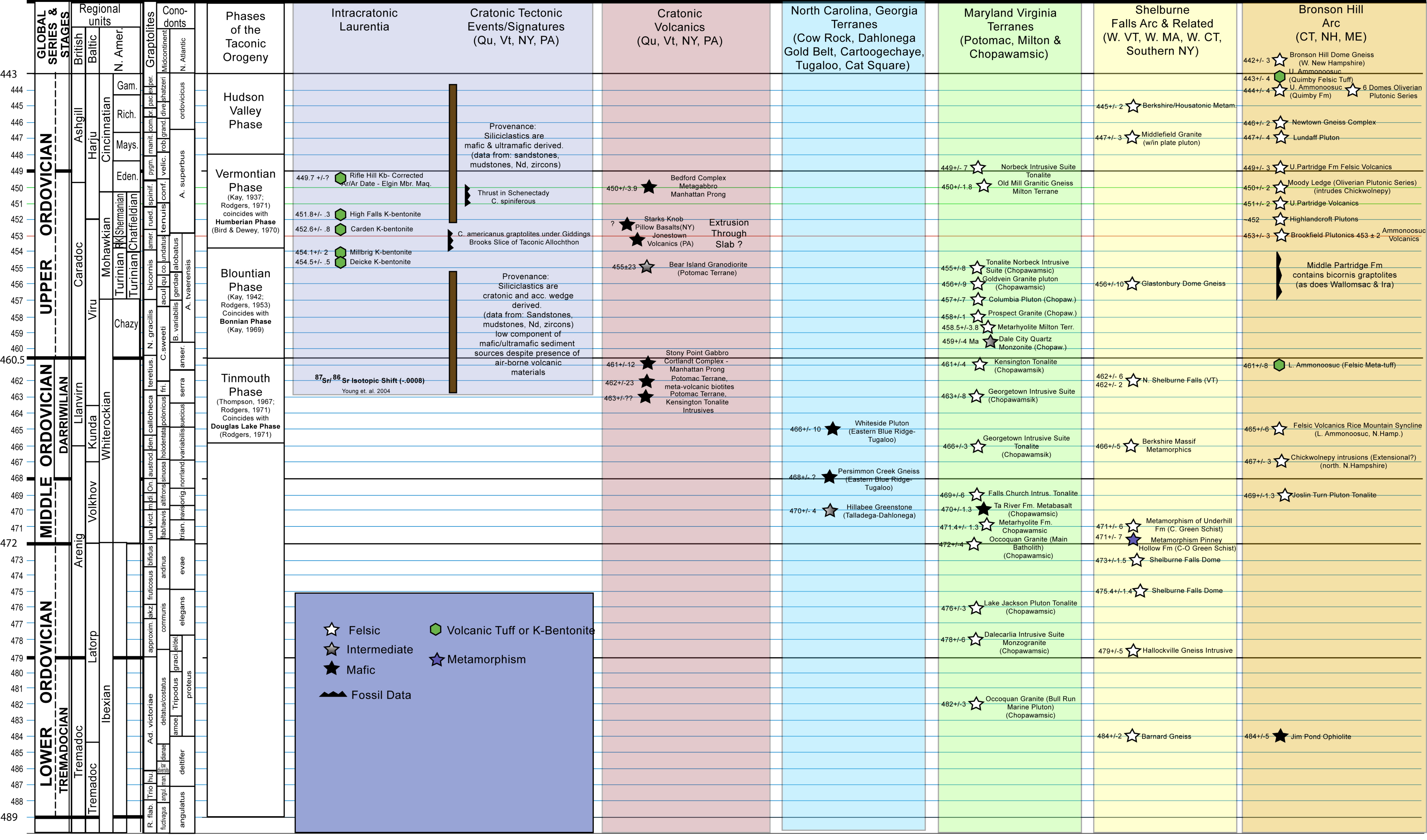
Figure 11: Geologic map of select features of the Appalachian region showing key "terranes" and allochthons that were involved directly in the Taconic Orogeny and/or were metamorphosed during the Ordovician with the exception of the Carolina Super Terrane, the Goochland Terrane and the Wilmington Complex (gray shades) which are likely Devonian. Compiled from multiple sources including: Bream et al., 2004; Gao et al., 2000; Hatcher, 1978; Hatcher, 2005; Holm, 2005; Horton et al., 1989; Karabinos et al., 1998.

were located in the Iapetus Ocean and were beginning to migrate towards the southeastern margin of Laurentia along a subduction boundary.

The distribution and thickness patterns of major K-bentonites including the Deicke and Millbrig beds seem to suggest a Carolina source for these volcanic ashes (Kolata et al., 1996). If this was indeed the source area for these volcanic ash beds, there is minimal direct evidence for the volcanic source area in that region today. Whether due to erosion and subsequent metamorphism, most preserved age-equivalent terranes in the Carolinas are metasedimentary or have volcanic signatures that suggest primitive “MORB-like” (Mid-Ocean Ridge Basalts) mafic signatures (Thomas et al., 2001; but see Meschter McDowell, et al, 2002). It is unlikely that these bentonites were derived from that immediate region. Nonetheless there are potential volcanic arc complexes. For instance intermediate volcanic signatures are found to the southwest in adjacent Georgia and Alabama (Persimmon Creek Gneiss, ~465 ma; Thomas, 2001), and to the north in Virginia (Chopawamsic and Baltimore region). Given the paleolatitudinal position, and wind patterns for equivalent modern latitudes, the paleowind directions for the Ordovician subtropical belt allow for source volcanoes along the entire cratonic margin from the Carolinas and Tennessee through offshore Vermont and potentially further. Nonetheless, as shown in **figure 12**, geochemical and field mapping data collectively suggest a complex collisional configuration of a series of island arc complexes, and associated terranes that interacted with the Laurentian margin beginning in the Late Ordovician.

Given 1) the complexity of field observations for Ordovician tectonism along the length of the orogen, 2) the apparent “scissor-like” collisional closing along eastern Laurentia (Ettensohn, 1991; 1994), 3) the apparent lack of a volcanic island arc in the vicinity of the Sevier foreland basin, and 4) the presence of various Blue Ridge/Piedmont belts with Laurentian

Figure 12: Geochronology of Ordovician tectonic, igneous, and metamorphic events in eastern North America as established in key areas of this study. Data as compiled from multiple sources including: Coler et al., (2000), Gillon, (2001), Hollocher & Robinson (2002), Karabinos et al, (1998), Karabinos & Hepburn (2001), Min et al. (2001), Samson et al. (1988), Young et al. (2004).



GLOBAL SERIES & STAGES
Regional units
British
Baltic
N. Amer.
Graptolites
Conodonts
Midcontinent
N. Atlantic
Phases of the Taconic Orogeny
Intracratonic Laurentia
Cratonic Tectonic Events/Signatures (Qu, Vt, NY, PA)
Cratonic Volcanics (Qu, Vt, NY, PA)
North Carolina, Georgia Terranes (Cow Rock, Dahlonega Gold Belt, Cartoogechaye, Tugaloo, Cat Square)
Maryland Virginia Terranes (Potomac, Milton & Chopawamsic)
Shelburne Falls Arc & Related (W. VT, W. MA, W. CT, Southern NY)
Bronson Hill Arc (CT, NH, ME)
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UPPER ORDOVICIAN
MIDDLE ORDOVICIAN
DARRIWILIAN
LOWER ORDOVICIAN
TREMADOCIAN
Ashgill
Caradoc
Llanvirn
Arenig
Tremadoc
Harju
Viru
Kunda
Volkhov
Latorp
Tremadoc
Cincinnatian
Mohawkian
Whiterockian
Ibexian
Gam.
Rich.
Mays.
Eden.
Chazy
Turinian
RK
Shermanian
Chatfieldian
A. superbus
A. tvaerensis
anser.
serra
variabilis
suecicus
norrland
orig.
navis
trian.
evae
elegans
proteus
deltifer
angulatus
ordovicicus
Hudson Valley Phase
Vermontian Phase
(Kay, 1937; Rodgers, 1971) coincides with Humberian Phase (Bird & Dewey, 1970)
Blountian Phase
(Kay, 1942; Rodgers, 1953) Coincides with Bonnian Phase (Kay, 1969)
Tinmouth Phase
(Thompson, 1967; Rodgers, 1971) Coincides with Douglas Lake Phase (Rodgers, 1971)
Provenance: Siliciclastics are mafic & ultramafic derived. (data from: sandstones, mudstones, Nd, zircons)
449.7 +/-? Rifle Hill Kb- Corrected Ar/Ar Date - Elgin Mbr. Maq.
Thrust in Schenectady C. spiniferous
451.8+/- .3 High Falls K-bentonite
452.6+/- .8 Carden K-bentonite
C. americanus graptolites under Giddings Brooks Slice of Taconic Allochthon
454.1+/- 2 Millbrig K-bentonite
454.5+/- .5 Deicke K-bentonite
Provenance: Siliciclastics are cratonic and acc. wedge derived. (data from: Sandstones, mudstones, Nd, zircons) low component of mafic/ultramafic sediment sources despite presence of air-borne volcanic materials
87Sr/86Sr Isotopic Shift (-.0008)
Young et. al. 2004
450+/-3.9 Bedford Complex Metagabbro Manhattan Prong
? Starks Knob Pillow Basalts(NY) Jonestown Volcanics (PA)
Extrusion Through Slab ?
455±23 Bear Island Granodiorite (Potomac Terrane)
461+/-12 Stony Point Gabbro Cortlandt Complex - Manhattan Prong
462+/-23 Potomac Terrane, meta-volcanic biotites
463+/-?? Potomac Terrane, Kensington Tonalite Intrusives
466+/- 10 Whiteside Pluton (Eastern Blue Ridge-Tugaloo)
468+/- ? Persimmon Creek Gneiss (Eastern Blue Ridge-Tugaloo)
470+/- 4 Hillabee Greenstone (Talladega-Dahlonega)
449+/- 7 Norbeck Intrusive Suite Tonalite
450+/-1.8 Old Mill Granitic Gneiss Milton Terrane
455+/-8 Tonalite Norbeck Intrusive Suite (Chopawamsic)
456+/-9 Goldvein Granite pluton (Chopawamsic)
457+/-7 Columbia Pluton (Chopaw.)
458+/-1 Prospect Granite (Chopaw.)
458.5+/-3.8 Metarhyolite Milton Terr.
459+/-4 Ma Dale City Quartz Monzonite (Chopaw.)
461+/-4 Kensington Tonalite (Chopawamsik)
463+/-8 Georgetown Intrusive Suite (Chopawamsik)
466+/-3 Georgetown Intrusive Suite Tonalite (Chopawamsik)
469+/-6 Falls Church Intrus. Tonalite
470+/-1.3 Ta River Fm. Metabasalt (Chopawamsic)
471.4+/- 1.3 Metarhyolite Fm. Chopawamsic
472+/-4 Occoquan Granite (Main Batholith) (Chopawamsic)
476+/-3 Lake Jackson Pluton Tonalite (Chopawamsic)
478+/-6 Dalecarlia Intrusive Suite Monzogranite (Chopawamsic)
482+/-3 Occoquan Granite (Bull Run Marine Pluton) (Chopawamsic)
445+/- 2 Berkshire/Housatonic Metam.
447+/- 3 Middlefield Granite (w/in plate pluton)
456+/-10 Glastonbury Dome Gneiss
462+/- 6 462+/- 2 N. Shelburne Falls (VT)
466+/-5 Berkshire Massif Metamorphics
471+/- 6 Metamorphism of Underhill Fm (C. Green Schist)
471+/- 7 Metamorphism Pinney Hollow Fm (C-O Green Schist)
473+/-1.5 Shelburne Falls Dome
475.4+/-1.4 Shelburne Falls Dome
479+/-5 Hallockville Gneiss Intrusive
484+/-2 Barnard Gneiss
442+/- 3 Bronson Hill Dome Gneiss (W. New Hampshire)
443+/- 4 U. Ammonoosuc (Quimby Felsic Tuff)
444+/- 4 U. Ammonoosuc (Quimby Fm)
6 Domes Oliverian Plutonic Series
446+/- 2 Newtown Gneiss Complex
447+/- 4 Lundaff Pluton
449+/- 3 U.Partridge Fm Felsic Volcanics
450+/- 2 Moody Ledge (Oliverian Plutonic Series) (intrudes Chickwolnepy)
451+/- 2 U.Partridge Volcanics
~452 Highlandcroft Plutons
453+/- 3 Brookfield Plutonics 453 ± 2 Ammonoosuc Volcanics
Middle Partridge Fm contains bicornis graptolites (as does Wallomsac & Ira)
461+/-8 L. Ammonoosuc (Felsic Meta-tuff)
465+/-6 Felsic Volcanics Rice Mountain Syncline (L. Ammonoosuc, N.Hamp.)
467+/- 3 Chickwolnepy intrusions (Extensional?) (north. N.Hampshire)
469+/-1.3 Joslin Turn Pluton Tonalite
484+/-5 Jim Pond Ophiolite
Felsic
Intermediate
Mafic
Fossil Data
Volcanic Tuff or K-Bentonite
Metamorphism

affinities, it is clear that the existing models for the Blountian-Vermontian collisions are in need of further evaluation. As postulated by Thomas and colleagues (2001), the absence of an arc in the Carolinas suggests that a small independent sliver or fragment of ocean floor was accreted prior to and distinct from the more substantial collisional episode associated with the more northerly tectophase. This model is illustrated in Ettensohn (2002b) where a left-lateral strike-slip boundary is postulated between the Blountian Highlands and the Carolina Terrane. Subsequently, Ettensohn (2002c) documented a slightly evolved model showing the major east-dipping subduction zone trending north-south and tangential to the Laurentian margin (modified after Scotese & McKerrow, 1991). This subduction zone was located well east of the Blountian region as the eastern boundary of an oceanic plate sliver located outboard of the Blountian Highlands, and the Blue Ridge Terranes. This sliver or wedge is depicted as tapering to closure to the north in the vicinity of the Pennsylvania Embayment where a large lowland gap between the Blountian (in the south) and the Taconian Highlands (in the north) is predicted. Although this model is thought provoking, the spatial-temporal resolution of this configuration needs to be addressed in order to evaluate the model further.

Cratonward of the continental margin, there are additional structural components that need to be mentioned (**figure 13**). These features have been suggested to have either formed or been reactivated due to far field tectonism inboard of collision belts. Many of these features show initial development during the Precambrian. From their sedimentary fills, it appears that many of these are related to initial rifting of Rodinia in the Neoproterozoic. At that time, the Laurentian region was fragmented at various times in the latest Precambrian and into the Early Cambrian. This rifting not only resulted in the opening of the Iapetus Ocean (main Iapetan rift is shown in figure 13 with the position of key transform segments), but the interior of the craton

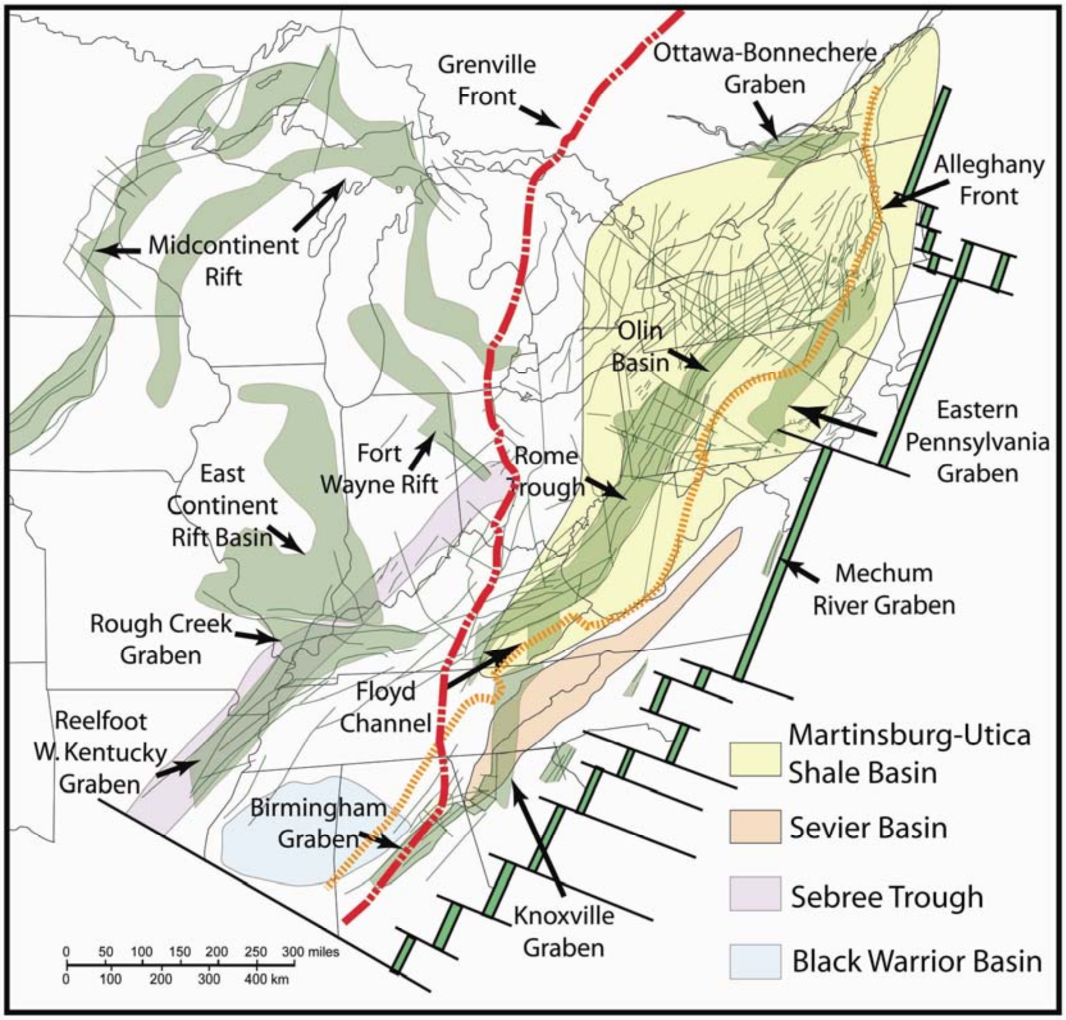


Figure 13: Key structural features of Eastern North America. Shown are the positions of most faults, fracture/joint intensification domains, and interior "rift basins" and grabens as recognized in the subsurface. In addition the approximate boundaries of each of the Upper Ordovician Taconic-aged sedimentary basins are shown. The position of the Laurentian cratonic margin during the Ordovician is also shown along with the boundary of the Alleghany fold and thrust and the Grenville basement Front.

was also structurally weakened in the process.

A number of fractures and normal faults initiated and resulted in the development of a number of northeast-southwest trending intracratonic basins and associated linear highs. As such, these were relatively parallel with the Laurentian margin. These features, including the Rome trough, the Reelfoot, Western Kentucky, Birmingham, Knoxville, and Eastern Pennsylvania grabens, the Floyd channel, and the mid-continent rift basin of Iowa, are bordered by faults and have been interpreted as major elements of the interior rift system. Evidence shows that they formed during the opening of the proto-Atlantic (Harris, 1978; Shumaker & Wilson, 1996; Gao et al., 2000). A number of northwest-southeast trending lineaments, oriented perpendicular to rift opening direction and including some within the well-studied Kentucky River Fault Zone, intersect these features and are shown to cause offsets in graben geometries and also disrupt patterns of basin fill (Gao et al., 2000).

Although formed in the earliest Paleozoic, these structures show evidence for periodic reactivation through the latest Cambrian and Early Ordovician – as is the case with the Rome Trough. Likewise, in the New York region similar structures in the Mohawk and St. Lawrence River Valley shows evidence of movement even during the Middle Ordovician (Fisher, 1977, Bechtel & Mehrtens 1995; Salad Hersi & Dix, 1999). In the mid-continent region, during the Ordovician and indeed during the Taconic Orogeny and its two tectophases, interactions of these structural weaknesses during the onset of lithospheric compression have been called upon to produce most of the region's paleotopographic features that influenced deposition, and paleoceanographic circulation patterns. Such features include the Sebree Trough, as well as the Jessamine Dome, the Nashville Dome, Waverly Arch, Tazewell Arch, as well as others.

Nonetheless, although a number of lithospheric flexural models have been proposed to explain the connections between orogenesis, foreland basin development, and intracratonic topographic expression, it is still unclear as to exactly how these features interact with the flexural wave as proposed in these models (Bayona & Thomas, 2003). Deep crustal research by Hawman and Phinney (1992 a, b) suggests that at least in Pennsylvania some of these highs (and associated lows) could have a deeper root associated with pools of mafic rock and high-grade metamorphics that underplated the crust in areas of the Great Valley. Dense rock seismic returns were interpreted to be mafic rocks in their investigation, but it is unclear from their study as to the source and timing of emplacement. If these mafics are not Precambrian in age, there is the possibility that these were emplaced during the Taconic Orogeny resulting in slight uplift of ancestral rift features either by load induced flexure or thermal buoyancy or combinations of both. The source of the mafic pools is enigmatic but perhaps these could be related to voluminous partial melting following subduction reversal as prescribed by Karabinos and colleagues (1998). Recent geochemical studies of Stark's Knob pillow basalts of New York, as well as the Jonestown Volcanics member in (*Corynoides americanus* graptolite zone) the Martinsburg Formation from the Hamburg Klippe suggest that these volcanics were likely extruded through carbonate platform sediments (Landing et al., 2003). If this is the case, there could be a connection between mafic under-plating and Taconic Orogenesis.

Nonetheless, it is clear that many topographic highs (commonly linear arches and semi-circular domes) and lows (usually linear troughs or semi-circular basins) became accentuated at different times during the Taconic Orogeny and influenced local sedimentation patterns to some degree. The subject of the following chapters, is to establish a robust chronostratigraphic framework documenting the spatial-temporal modification of rocks in proximal (intracratonic) and distal (cratonic marginal) settings in order to investigate more confidently the timing and relative impact of Taconic orogenesis on the architecture of the GACB through its final demise.

CHAPTER 2: Methods and Correlation Tools for Unraveling the Stratigraphy of the Great American Carbonate Bank

INTRODUCTION AND PURPOSE:

Wide-spread carbonate production, and the accumulation of substantial thicknesses of limestone, is only possible under a specific range of environmental conditions that are persistent for a limited time (Bosscher & Schlager, 1993; Webb, 2001). In most cases, large carbonate platforms (1000's of square kilometers in area) are deposited during: 1) periods of rising sea-level when shallow water environments become pervasive over flat platforms devoid of siliciclastic sediments (occur during transitional climates like those of the modern Bahamian platform), 2) global sea-level highstands associated with climate maxima and periods of rapid seafloor spreading (large epicontinental seas are developed such as the one that covered much of proto-North America during the Late Cambrian to Early Ordovician), or 3) during the transition from passive tectonic margins into actively subsiding foreland basin settings associated with collisional tectonics.

An increasing number of studies have shown that the Late Ordovician Great American Carbonate Bank (GACB) of eastern Laurentia was deposited during 1) the early phases of the Tippecanoe transgression, 2) the onset of the Taconic Orogeny, 3) a major climate greenhouse maxima, and 4) a period that witnessed a substantial burst of origination and radiation of many marine taxa. Moreover, this work has shown that the final demise of the GACB was coincident with: 1) later tectophases of the Taconic orogeny signified by rapid subsidence of portions of the platform, 2) delivery of significant volumes of siliciclastic sediment, 3) an interval of cooling and/or periodic climate volatility, and 4) a period of reduced origination and increased extinction frequency leading up to the end Ordovician. Clearly, the sequence of events leading up to the demise of the GACB appears to be intimately linked to sea-level, climate, and tectonic change;

however, it is not entirely clear as to the timing and relative impact of tectonics, sea-level fluctuations, and/or the nature of climate change across the platform. The lack of a detailed, high-resolution stratigraphic synthesis for the GACB region is the primary hurdle for making inroads into these problems.

Hence, the purpose of this study is to review and integrate herein a number of correlation tools, exclusive of sequence stratigraphy, that have been used to establish local correlations. Here these tools are synthesized between important outcrop regions of the GACB in order to construct a regionally applicable stratigraphic framework. Some stratigraphic methods, when compared to other methodologies, produce inconsistent correlations within certain stratigraphic intervals (i.e. the Black River-Trenton boundary interval). In all cases particular attention is paid to use one correlation technique against another for testing the accuracy of time correlations and to build as consistent a stratigraphic synthesis as possible. Thus ultimately the goal of this discussion is to establish the most reasonable and parsimonious model for this portion of the upper Ordovician that can be subdivided by subsequent sequence stratigraphic analysis into calibrated time-slices (see chapter 7) in order to evaluate the timing and scale of environmental change across significant portions of the GACB (see chapter 8).

STUDY AREA & STRATIGRAPHIC METHODS

This study focuses on three primary outcrop regions across eastern Laurentia: 1) the Chazy-Black River and Trenton Group type sections of New York, Vermont, and Ontario, 2) the equivalent succession exposed in central and south-central Pennsylvania in the Ridge and Valley Province, and 3) the High Bridge Group and Lexington Formations of central and northern Kentucky and southwestern Ohio. This study has worked to integrate both outcrop and subcrop studies from all of these regions where possible. In addition, other important and critical areas of eastern North America are also mentioned where necessary to establish a broader framework of correlations. In these cases, stratigraphic analyses and correlations are established based primarily on literature review; however, in some cases these are supplemented by field studies by the author in order to gain a personal understanding of former stratigraphic studies. Included in the following discussion of stratigraphic methodologies are: 1) a review and integrated synthesis of traditional biostratigraphic data from the Chazy, Black River and Trenton Groups, and 2) a review and integration of a growing list of event stratigraphic horizons including K-bentonites, as well as chemostratigraphic events. An additional chapter (chapter 6) has also been dedicated to investigation of unique lithostratigraphic events that appear to have relatively widespread signatures across portions of the GACB. Moreover, the latter are proving to be significant and important for identifying potential climatic and tectonic linkages. The implication of these correlations will be discussed in a subsequent chapter (chapter 8).

BIOSTRATIGRAPHY

Previous biostratigraphic analyses of both macro and microfauna, including conodonts and graptolites, have provided initial constraints for stratigraphic correlation between the outcrop regions investigated in this study. Substantial progress has been made by many previous authors to help establish a reliable chronostratigraphy for the Mohawkian Series of North America. **Figure 1** represents a compendium of chronostratigraphic data for the Upper Ordovician of the eastern United States. In this diagram, the distribution of key graptolite and conodont taxa are shown relative to lithostratigraphy and Ordovician Time Slices as defined by Webby and colleagues (2004). Radiometric age constraints and time-rock nomenclature (series and stages)

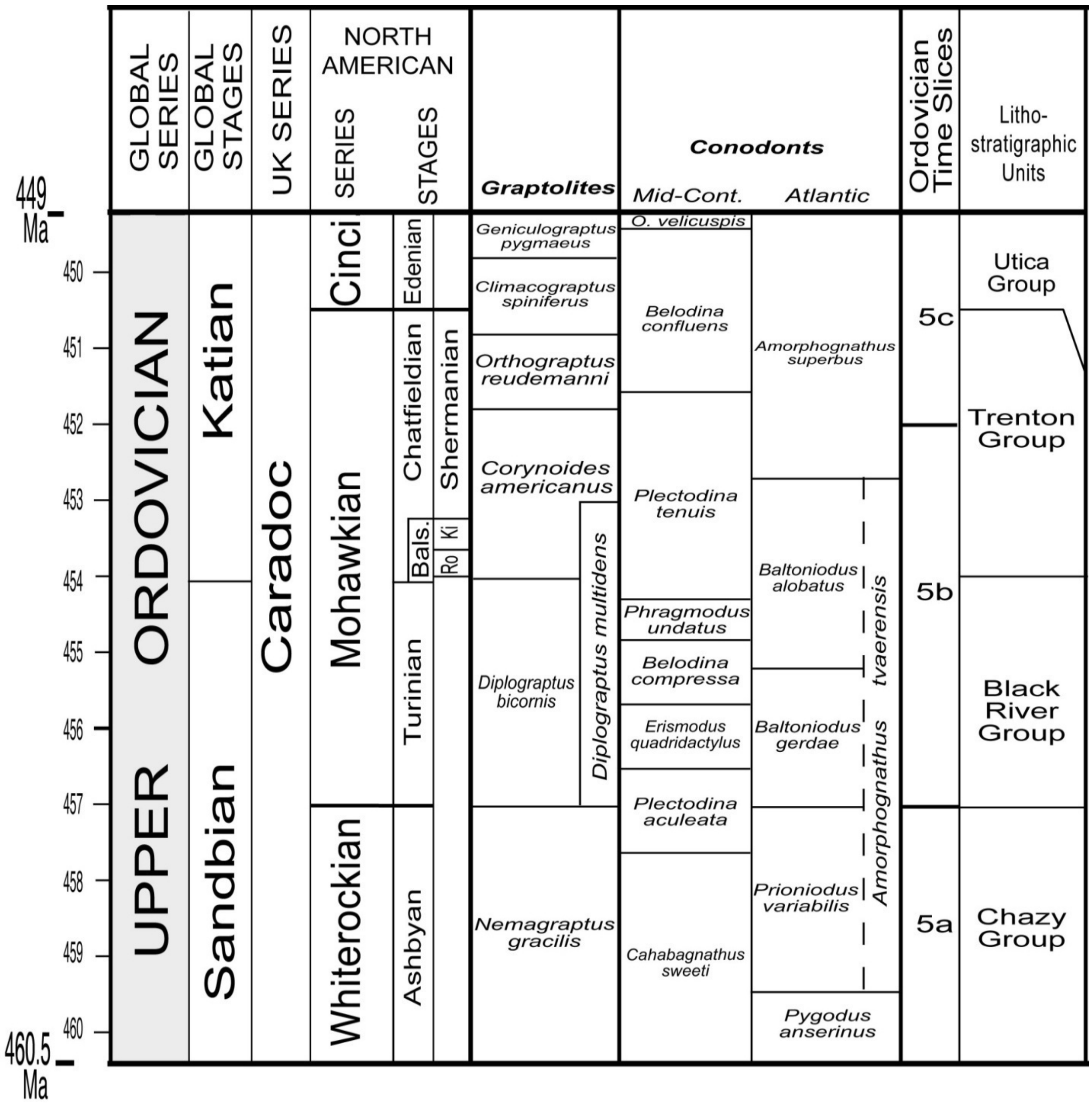


Figure 1: Upper Ordovician conodont and graptolite biostratigraphy for the Chazy, Black River, and Trenton Groups. Biostratigraphy, and time slices are applied after Webby et al., 2004 and reflect data from Mitchell & Bergstrom (1991) for graptolite occurrences, and from Leslie, (2000) for conodont biostratigraphy. Abbreviations: Bals.= Proposed Balsamian substage (Barta et al., in press) Ro.= Rocklandian, and Ki. = Kirkfieldian.

are from the same publication. For consistency to older literature, older stage and substage classifications are included. In most cases, the boundaries of graptolite and conodont biozones do not correspond directly with time series and stage boundaries. The relative position of key graptolite and conodont species within each representative sub-unit of the Middle to Upper Ordovician are shown and has been refined and applied here from previous studies.

Graptolite and Conodont Biostratigraphy

Despite the challenges in some intervals for discovery and identification of graptolites, they are nonetheless very important especially in siliciclastic facies. In general, the key graptolite taxa used for dating in the Chazy, Black River, and Trenton groups and their equivalents include: *Nemagraptus gracilis*, *Diplograptus bicornis*, and *Corynoides americanus* (which overlaps with *Diplograptus multidens), Orthograptus ruedemanni,* and *Climacograptus spiniferus.* The overlying Cincinnati Group equivalents contain the upper limit of the *Climacograptus spiniferus* and the basal *Geniculograptus pygmaeus* boundary (Mitchell & Bergström, 1991). As this interval is dominated by black shale deposition – these graptolites are usually very abundant and easily identified.

Mitchell and Bergström (1991) used the position of graptolite zonal boundaries in the type region of the Trenton Limestone and the type region of the Edenian Kope Formation (lower Cincinnatian) to establish correlations between these localities. Their study focused on the middle and upper Trenton (Denley to Steuben Formations) and Utica Shale. Unfortunately the distribution of graptolites in the lower, carbonate-dominated Trenton Limestone is poorly constrained. The lack of preservation of graptolites in the shallow, well-oxygenated depositional settings of much of the Trenton and Lexington limestones (of Ohio-Kentucky) limits the success of biostratigraphic assessments within these strata. Therefore, graptolite biozonation is supplemented by considerations from conodont biostratigraphy for this interval.

Also in their study, Mitchell and Bergström (1991) locate the approximate position of the *Amorphognathus tvaerensis / Amorphognathus superbus* conodont biozone boundary at the base of the Poland Member of the Denley Formation. This correlates with an interval in the middle of the *C. americanus* graptolite zone, below the base of the Dolgeville Formation along the boundary of the Taconic foredeep basin (Mitchell et. al, 1994; Jacobi & Mitchell, 2002). This puts the position of this conodont biozone boundary within the Shermanian Stage. Thus the biostratigraphy of Webby and colleagues (2004) is slightly adjusted in this regard – that is the position of the *tvaerensis-superbus* boundary is lowered to reflect these former assessments.

Conodont biostratigraphy for this interval is also complicated because conodonts appear to be strongly tied to facies and depth gradients. Moreover the type Black River –Trenton region spans two major conodont biogeographic provinces. These two biogeographic provinces include the Mid-Continent Province and the North Atlantic Province. The key conodont taxa recognized in the Mid-Continent Province include: *Cahabagnathus sweeti*, *Plectodina aculeata* (which straddles the Ashbyan-Turinian stage boundary), *Erismodus quadridactylus*, *Belodina. compressa*, *Phragmodus undatus* (straddles the Turinian-Chatfieldian boundary), *Plectodina tenuis,* and *Belodina confluens* (which extends past the Chatfieldian and into the Edenian stage of the Cincinnatian Series).

Key taxa recognized in the North Atlantic Province include; *Pygodus anserinus* (extends across the Middle-Upper Ordovician boundary), as well as the three characteristic subzones found in the *Amorphognathus tvaerensis* zone (*Prioniodus variabilis*, *Baltoniodus gerdae*, and *Baltoniodus alobatus*). As mentioned the last North Atlantic conodont zone includes *Amorphognathus superbus* which extends upward to the top of the Maysvillian (Sweet, 1984; Mitchell and Bergström, 1991; Brett and Baird, 2002). Sweet (1984) established the initial relative distribution of conodont forms across the United States (Nevada to New York). Sweet plotted the distribution of conodonts from 61 geographic localities, and created a 477 meter-thick composite stratigraphic section (CSS) using graphic correlation techniques. Sweet divided the CSS into 80 individual 6 meter-thick intervals, and into a series of chronozones based on first appearance of key conodont species from the mid-continent province for the Mohawkian and Cincinnatian Series (six conodont chronozones for the Mohawkian). This initial analysis established three chronozones for the Black River interval and three for the Trenton. Originally, the lower 3 chronozones included *P. aculeata* (used to define the base of the Mohawkian), *E. quadridactylus* and *B. compressa.*

The top of the Turinian – base of Rocklandian was originally defined by the first appearance of *Phragmodus undatus*. The two subsequent upper Mohawkian chronozones include that of *P. tenuis* and *B. confluens* which ranged upward into the Cincinnatian. Subsequently, *P. undatus* has been found well down in the Black River Group well below the Turinian top (Leslie, 2000) and therefore the chronostratigraphy provided by Webby and colleagues (2004) also adapted and shown in **figure 1** above reflects this change.

Unfortunately, Mitchell and Bergström (1991), and Sweet (1984) used two different conodont biochronologies that still need some reconciliation. In the case of Mitchell and Bergström (1991), the *A. tvaerensis* and *A. superbus* conodont biozonation is established from studies of faunas found in the North Atlantic conodont province, while that of Sweet (1984) is focused on faunas from the midcontinent. Unfortunately, in most cases the two faunas do not overlap in their geographic ranges. The different biostratigraphic zonations result from the general lack of North Atlantic species in most midcontinent localities and vice versa. In the case of the type Mohawkian interval in central New York, the conodont faunas are generally allied with those of the North Atlantic as is reflected by Mitchell and Bergström (1991). Nonetheless Sweet's (1984) graphic correlation, based on midcontinent faunas, has some overlap with sections from central New York and southern Ontario and therefore there are some cross-linkages.

Leslie (2000) significantly helped to establish these cross-linkages between the two conodont provinces and has additionally refined the first appearances (and therefore conodont biozones) of a range of key taxa. As a result of this work, Leslie (2000) has not only adjusted the base of the *P. undatus* biozone to a level lower in the Black River Group, but he has also lowered the *P. tenuis* first appearance. Leslie now infers the base of the *P. tenuis* chronozone to a level above the Millbrig K-bentonite (~2.9 m above it) in the basal Hermitage Formation of Tennessee. This position allows its association with the Millbrig K-bentonite and puts the chronozone boundary in the vicinity of the basal Chatfieldian and, based on this position, places it relatively near the basal Rocklandian sub-stage boundary.

Recent re-investigation of faunal lists of the Ottawa Valley published by Barnes (1967) denotes the occurrence of *Ozarkodina tenuis* Branson and Mehl from the Chaumont Formation of that region and records from his unpublished Ph.D. dissertation (1964) report the occurrence of this form in the underlying Lowville Formation. Recent personal communication (2007) with Dr. Barnes indicates that *O. tenuis* of his 1964 work should be updated to the current *P. tenuis* form. Thus the identification of *P. tenuis* in the Hermitage of Tennessee although significantly lower than previously assessed may still not be the lowest occurrence of this form if the occurrence of *O. tenuis* in the Lowville is accurate. The occurrence in the Lowville would represent the first occurrence and significantly depress the position of the *P. tenuis* zonal boundary to a position not distinctly different from that of the presently recognized *P. undatus* base.

Nonetheless, the *P. tenuis* conodont chronozone needs to be better defined for sections in NY, PA, KY, and elsewhere if it is to be used reliably for establishing correlation. To date formally recognized lowest occurrences in these latter regions (as per Sweet, 1984) correlate with positions in the middle Trenton in New York, the upper-Salona-base Coburn in Pennsylvania, and the middle Lexington Group in Kentucky (Richardson & Bergström, 2003). In all cases each of these *P. tenuis* occurrences is located well above the base of type Trenton lithologies, above the base of the type Rocklandian based on other independent correlations, and likely well above the true first occurrence of the form. Moreover, Richardson and Bergström (2003), in the analysis of conodonts from outcrops and cores in the Kentucky- Ohio-Indiana region show the lowest occurrence of *P. tenuis* to be some 50 to 70 feet above the base of the Lexington Limestone, where they recognize the lowest occurrence of this conodont in the upper Grier Formation –the Perryville of this study. The lowest portions of the Lexington Group – which are well correlated with the Hermitage of Tennessee, do not yet show *P. tenuis* in the base of the Lexington although its position is inferred by association with the Hermitage. Clearly the first occurrence of *P. tenuis* has not been identified in most all outcrop regions to a level at or below (?) the Millbrig K-bentonite.

Barnes (pers. com., 2007) suggests that *P. tenuis* (synonymous with his *O. tenuis*) is strongly tied to shallow facies not commonly observed in typical Trenton strata – therefore its recognition may not be possible in many regions where depositional facies are substantially deeper. Therefore it is clear that conodont biostratigraphy, although significantly advanced over earlier analysis, still requires additional consideration and refinement. Most importantly, the association of *P. tenuis* with the basal Hermitage of Tennessee and with the Chaumont Formation helps to establish a direct tie line for correlation of the type Rocklandian outside of NY-Ontario. If the Chaumont Formation underlies type Rocklandian, then clearly the Curdsville Limestone at the base of the Hermitage of Tennessee and Kentucky is older than Rocklandian and not younger than it as previously assessed. Thus the age of the Curdsville Limestone and underlying Tyrone Limestone cannot be Kirkfieldian and Rocklandian, respectively. Moreover, the first occurrence of *P. undatus* in the type region, which was used to define the base of Rocklandian, can no longer be used to define this chronostratigraphic position elsewhere because it has been found well down in strata below the Millbrig K-bentonite. Clearly additional analysis is needed especially in the region of New York, Pennsylvania, and Kentucky –Ohio where the position of these conodonts needs to be defined confidently.

Given the foundation of these previous workers and as shown in **figure 1** the lower Upper Ordovician can thus be divided into a number of different biozones and chronozones established on the basis of graphic correlation techniques. According to refinement of the Ordovician time-scale and new radiometric dates, a refined absolute chronology is helping to identify duration for each of these chronozones (Webby et al., 2004). Each chronozone thus ranges from 1 to slightly more than 2 million years. Webby and colleagues also identified 3 distinct time slices for the intervals considered in this study. Their Time slice 5a coincides approximately with Chazy Group (Ashbyan stage), 5b coincides with the Black River Group through middle Trenton Group (Mid-Shermanian stage), and 5c includes upper Trenton through Cincinnati groups.

<u>Macrofaunal Biostratigraphy</u>

The use of macrofauna for biostratigraphic purposes within the Ordovician has perhaps been one of the most disconcerting dilemmas for paleontologists working on this interval. Due in part to the rapid diversification of faunas in the Ordovician radiation, this interval has been a major interval for paleontologic research. At the same time, as environmental change also appears to have been very substantial most macrofaunal taxa are benthic forms and therefore their occurrence is facies controlled. Thus, their use as index fossils and the establishment of distinctive chronozones has been a challenge. Nonetheless, there are a number of taxa throughout the latest Whiterockian and Mohawkian interval that do aid in refinement of the existing biostratigraphic framework. Moreover although their first appearance and last appearances may be challenging to ascertain, in most cases their appearance in acme zones or epibole intervals are recognized and are often tied with cyclic facies changes that provide opportunities for higher resolution correlations.

A variety of taxa (corals, brachiopods, bryozoans, mollusks, trilobites, ostracods, echinoderms, etc.) have been used historically to define individual sub-units throughout much of New York State and Ontario. Through the mid 1900s, researchers such as White (1896); Prosser and Cummings (1896); Raymond (1903); Cushing and others (1910) and Kay (1937; 1943; 1968) as well as others, investigated and reported the biostratigraphic zonation of the type Mohawkian limestones. It was based on these studies that the interval was divided up into various sub-stages for correlation outside of the type region and it was these initial studies that were applied outside of New York to establish the prevalent use of the New York nomenclature through much of the Appalachian Basin province as well as much of the mid-continent region.

Since that time, there has been much confusion in the literature as to the application of stratigraphic terminology outside of the New York/Ontario/type regions. This confusion relates to both rock terms and time-rock terms – owing especially to the inability to recognize the full suite of macrofauna and lithologies present in the type regions. This issue has been to some degree superseded by conodont and graptolite biostratigraphy – although as mentioned, even these systems have their difficulties.

Thus without significant biostratigraphic control from graptolites or conodonts, the Turinian of New York State is generally defined based on the prevalence of fine-grained "birdseye" limestone containing abundant specimens of the tabulate coral *Tetradium cellulosum* and related taxa. By extrapolation with bracketing units, the Turinian basal boundary is now placed in the approximate position of the first appearance of *D. bicornis* (graptolite) and the first appearance of *B. gerdae* (conodont). The Turinian extends upward to the base of the Rocklandian Stage which was defined on the basis of brachiopod assemblages including the appearance of *Hesperorthis tricernaria, Triplecia cuspidata, and Doleroides ottawanus* and a number of dalmanellid brachiopods. The basal Kirkfieldian stage is defined on the basis of an acme zone of brachiopods called *Parastrophina hemiplicata* as well as an abundant assemblage of unique echinoderm taxa. These echinoderm taxa are especially prevalent in the Hull Limestone of the Ottawa region, and the type Kirkfield Limestone of the Lake Simcoe district and from a number of other time equivalent localities.

In contrast to the basal Trenton intervals, much of the middle to upper Trenton limestones (dominated by the Sugar River, Denley, and lower Rust formations) are typified by a variety of taxa that taken collectively can be considered a typical "Trenton assemblage." Thus, Kay (1968) referred to the entire middle Trenton using Shermanian as a time-rock term. Key taxa of this interval included the bryozoan *Prasopora simulatrix* (and other similar forms), a range of trilobite taxa including *Ceraurus pleurexanthemus, Isotelus gigas, and Flexicalymene senaria,* as well as a wide variety of gastropod and cephalopod mollusks. Although he only included formations up to the top of the Denley within his Shermanian stage; (for rocks above the Denley he applied the term Cobourgian), the remainder of the Upper Trenton is now all considered Shermanian in age. The development of the *Rafinesquina deltoidea* zone within the Steuben Formation (uppermost Trenton) was considered a unique faunal zone and Kay successfully correlated it with outcrops in Ontario where he obtained the name. *Rafinesquina deltoidea* first appears in the upper Rust Formation and expands dramatically in the Steuben and this appears to be the case in many time equivalent units. This expansion is also coincident with the *C. spiniferus* graptolite zone which helps to identify this event.

Based on the occurrence of these faunal zones, Kay (1937; 1943; 1968) established a number of time-rock terms to emphasize the distinction between biostratigraphic and lithostratigraphic terminologies (**figure 2**). Kay (1968) considered the entire Trenton Limestone

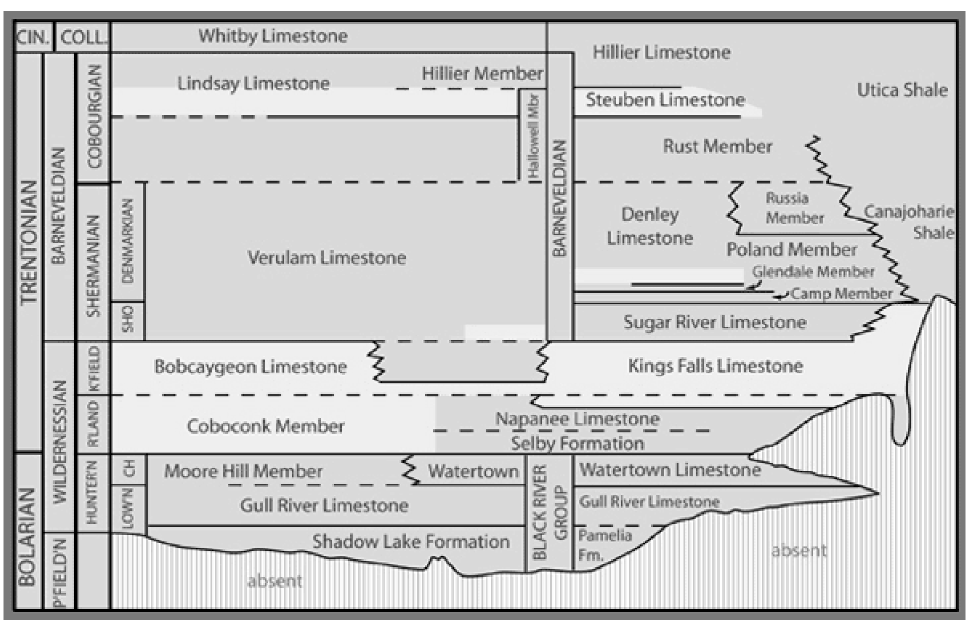


Figure 2: Figure modified after Kay, (1968) showing lithostratigraphic units of the Trenton for the Ontario, Canada to central New York State region. Also shown are the range of time-rock terms that he applied to denote chronozones most of which are no longer recognized. As shown, Kay preferred to consider the Trenton in terms of a time-rock concept (Trentonian) rather than as a lithologic term.

interval as belonging to one series which he termed Trentonian (although Trenton was used in the literature primarily as a rock-term). He also invented the terms Wildernessian, and Barneveldian as stages. In this same study, he also identified four time-rock stages (Rocklandian, Kirkfieldian, Shermanian, and Cobourgian), and an additional series of sub-stages within the Shermanian (Shorehamian and Denmarkian).

The development of Kay's stratigraphic framework was the result of many years of research and integration of multiple correlation methods including lithostratigraphy, event stratigraphy (K-bentonite correlation), as well as biostratigraphy. These assessments, although useable in the type region, were incredibly nomenclature rich and the application of this complex of time-rock terms outside of the region has been difficult. This study seeks to improve on this biostratigraphic assessment through the use of multiple correlation strategies so that it can be applied over wide regions of North America.

Coenocorrelation Studies

As mentioned, often the recognition and application of single taxon based biozones in correlation have been very difficult. Nonetheless, the "typical Trenton" faunas, although of poor use in biozonation, have proven useful as a tool for showing distinct environmental gradients through analysis of abundance patterns. In several studies it has been shown that individual taxa can be grouped into assemblages of fossils that often occur together and tend to remain in association (Cameron, 1968; Titus, 1977; Gildner, 1990). Cisne and Rabe (1978) used gradient analysis and coenocorrelation (community correlation) to generate correlations for the Trenton Limestones. They were able to document and demonstrate trends in the vertical and lateral changes in community composition as observed by previous authors.

Cisne and others (1982) used the position of K-bentonite (volcanic ash) horizons in localities from Trenton Falls eastward into the Mohawk Valley to constrain individual stratigraphic intervals for sampling fossils. They were able to follow these intervals from the medial Trenton at Trenton Falls down slope into deeper water facies of the equivalent shales (now called Flat Creek shales). Along this single time-space transect (~60 km), the relative abundance values for each taxon observed changed sequentially and statistically indicated a significant shift in an unknown parameter. These authors attributed the gradient change to water depth increase from shallow to deep from the Trenton Shelf into the Taconic foredeep basin. In their studies, Cisne and colleagues have shown that the development of faunal assemblages can change in time and space. In their estimation, the change in relative abundance of individual taxa is the result of environmental change resulting from relative sea-level fluctuation and the migration of the Taconic Foreland Basin into the Mohawk Valley region.

Gildner (2003) used another statistical analysis similar to those used in gradient analysis and coenocorrelation. He applied ordination techniques to try to establish environmental gradients using relative abundances of a variety of taxa. In this scenario, changes in relative abundances of various taxa reflect changes in any number of things: temperature, salinity, turbidity, water depth, etc. Similar to the reciprocal averaging technique used by Cisne, Gildner used a statistical treatment called detrended correspondence analysis (DCA), to explain the variation between samples, between time-slices, and between localities (**figure 3**). In the case of the Mohawk Valley the most prominent gradient had DCA scores ranging from low (to the west and in the basal Trenton) to high (to the east and high in the Utica).

In similar fashion to that of Cisne and Rabe (1978), Gildner's data suggest that the shift in faunal assemblages is tied to water-depth change such that Trenton Falls shows the shallowest

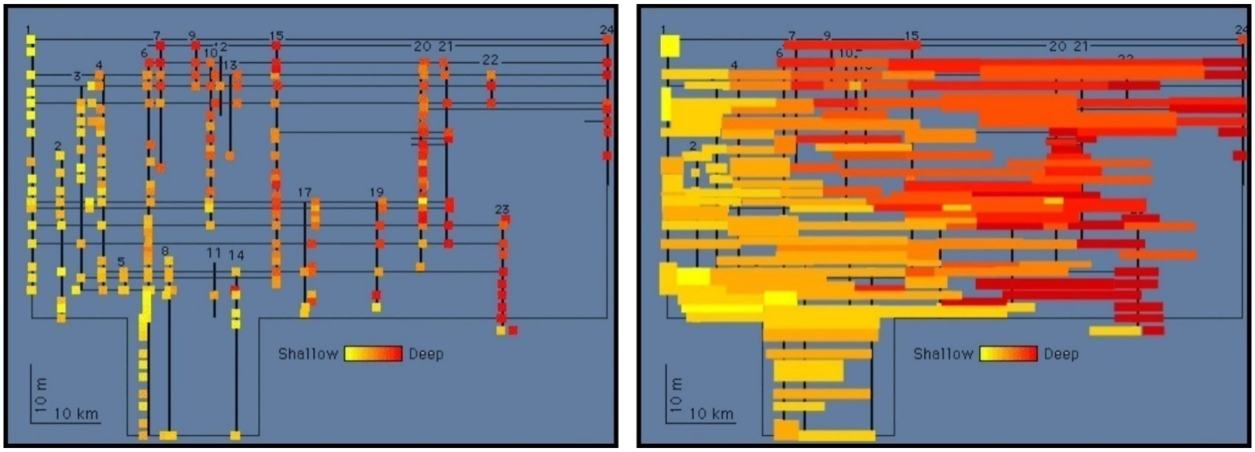


Figure 3: Detrended Correspondence Analysis Ordination Scores of faunal assemblages for a ~100 km transect from Trenton Falls (locality 1) southeastward along the Mohawk Valley to the area of Van Wie Creek east of Canajoharie. The diagrams have been color coded based on ordination scores that have been attributed to water depth and/or prominent lithologies. Clearly observed in the diagrams is the general east to west progression of deeper water facies associated with the transition from limestones to shales. In addition, the model to the right shows some of the higher-order signals observed in the data that show small-scale progradation and retrogradation of facies. Redrawn and modified after Gildner (2003) data available at http://zircon.union.edu/gildner/WD.html Localities are: 1. Trenton Falls, 2. Gravesville, 3. Mill Creek, 4.Poland, 5. Rathbun Brook, 6. Shedd's Brook, 7. Buttermilk Creek, 8. Farber Lane Brook, 9. Stony Creek, 10. Miller Road Creek, 11. North Creek, 12. Gun Club Road, 13. New York Thruway West, 14. New York Thruway East, 15. New York Route 5S, 16. West Crum Creek composite, 17. Dolgeville Dam, 18. East Canada and East Crum Creeks, 19. Mother Creek, 20 Caroga Creek, 21. Canajoharie Creek, 22. Flat Creek, 23. Currytown Road, 24. Van Wie Creek

facies and taxa overall, while deep water facies exist generally high in the section and to the east. Upon closer inspection there are also a number of higher-order patterns. This data clearly shows a series of higher-order, cyclical patterns in relative faunal abundances. This has also been attributed to the smaller-scale cycles found in the Trenton. These are likely deepening and shallowing phases with taxa tracking their preferred environments as has been suggested by Titus (1986). Nonetheless although these trends likely reflect relative water depth change, it is possible that these could reflect differences in turbidity, salinity, temperature driven by climatic oscillation as well as sea-level change.

Epibole Biostratigraphy

In this study, existing conodont and graptolite biostratigraphic frameworks (described previously) have been integrated with macrofaunal fossil occurrences ranging from the Chazy through the Trenton (**figure 4**). Each taxonomic group shown in figure 4 is a key or representative genus recognized over wide regions of North American during this time. In most

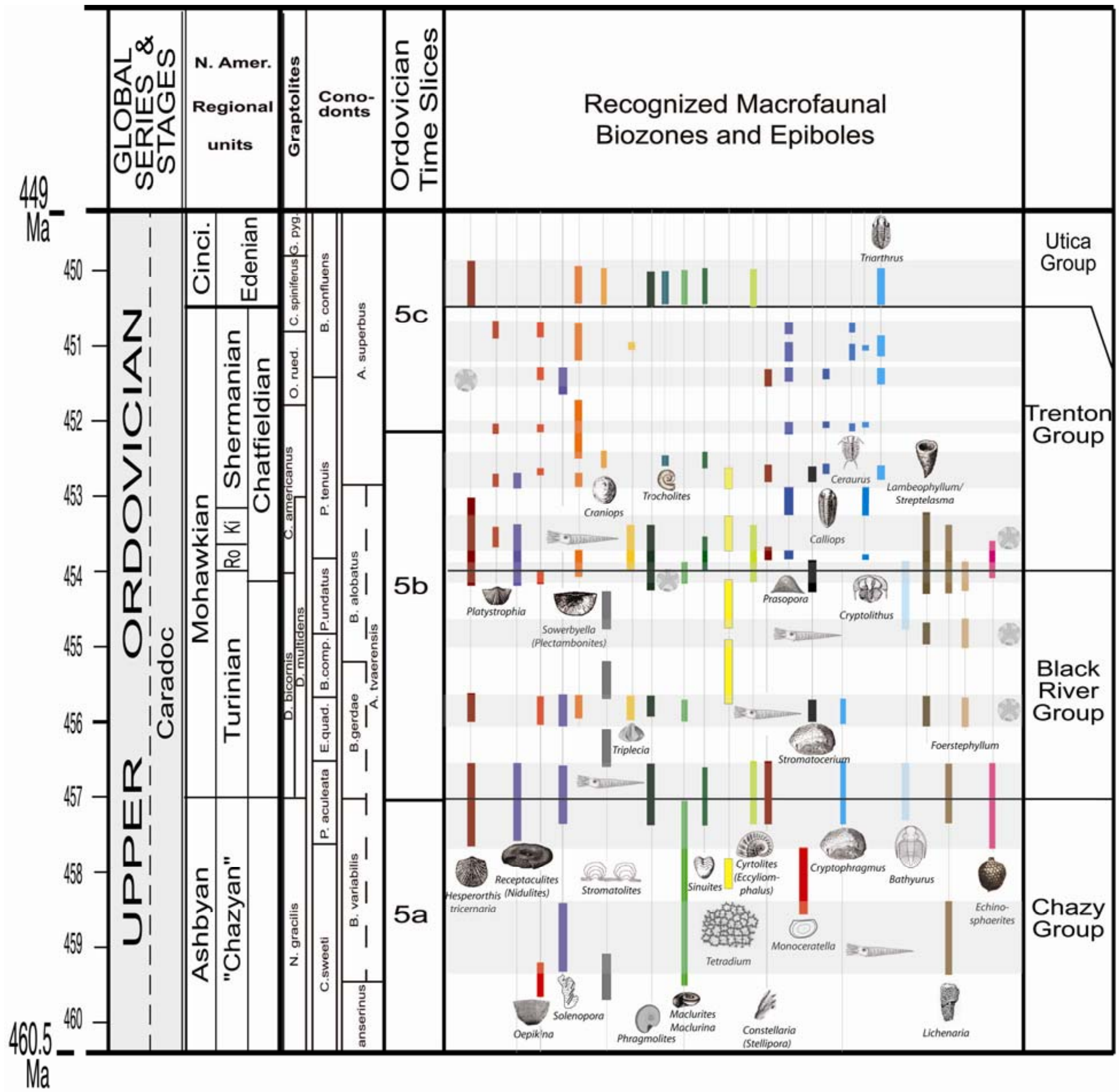


Figure 4: Characteristic and commonly recognized taxa from the Chazy, Black River, and Trenton groups from eastern North America. Range of each group (genus) is based on occurrence of one or more sister species, data is integrated from multiple published references and supplemented by data from the Paleobiology Database.

cases, these taxa represent individual species, while others are occurrences of like species included within the same genera. As is the case with the Gildner (2003) and Cisne and Rabe (1978) data, it is readily apparent that the recurrence zones of groups of taxa are not just limited to the Trenton. In fact, there are several key recurrence zones in the Chazy and Black River groups as well. What is also interesting is that, at least several of the taxa shown, tend to appear at roughly the same time over wide areas of the platform and then depart at nearly the same time, only to recur at a later time. Thus, the distribution and recurrence of taxa in the Chazy-Black River interval, is likely tied to sea-level oscillation and the development of open marine circulation in the GACB, while their extirpation is likely related to shallowing and restriction of the normal marine circulation. In most cases, this is supported simply by the fact that the intervening rock units tend to be dominated by restricted peritidal facies with stromatolitic development, evaporitic textures, and endemic low diversity faunas including small *Stictopora* bryozoans and leperditian ostracods.

In this analysis, several key taxa have been used to establish a baseline for this framework upon which additional taxonomic occurrences have been plotted. Most fossil occurrences have been tabulated using the constraints of initial conodont and graptolite biostratigraphic assessments. Once plotted graphically, individual stratigraphic units and their faunal lists were compared. Where unique taxa occurred on several faunal lists from different outcrop regions, they were plotted to help further refine the zonation as shown. As mentioned, most taxa were lumped into a single genus, although in several cases, individual species were characteristic of only one recurrence interval. These instances provide significant confidence in establishing correlations.

For example *Maclurites magnus* Lesueur and a closely related form *M. nitidus* (Rohr, 1980), interpreted to be epifaunal suspension feeders, are known from the type Chazy, the basal Stone's River Group of the Nashville Dome (Murfreesboro, Pierce, and lowest Ridley formations), the Whitesburg, Lincolnshire, and Lenoir formations of eastern Tennessee and southwestern Virginia. It is also known from the Mingan Islands north of Anticosti Island in Quebec in rocks considered to be the equivalent to the Chazy group of the Lake Champlain region. There is one recorded appearance of *M. magnus* in the Nashville Dome in a higher interval in the Lebanon Limestone (Wilson, 1949) where it is likely associated with a second incursion. Thus, the occurrence of this form helps to constrain the position of the type Chazy, together with *N. gracilis*, and *C. sweeti*, across the eastern margin of the GACB. Given this distribution, the *M. magnus* zone terminates at or just above the top of the *N. gracilis* graptolite zone, with only a limited reappearance in the *E. quadridactylus* zone in the Lebanon. Unfortunately, *M. magnus* is not recognized in other units in localities to the center of the GACB, including the Cincinnati Arch region, or the Upper Mississippi Valley where time equivalent strata lack the characteristic facies. However another macluritid form is recognized in the Camp Nelson Limestone of central Kentucky. In this case, *Maclurites bigsbyi* has been recognized on some bedding planes (Jillson, 1931), however initial investigation of the plates suggests this form could be *M. magnus* and is thus likely of similar to the occurrence in the Lebanon Limestone of Tennessee.

A subsequent incursion at the top of the Black River Group and in the basal Trenton Group demonstrates the return of a closely related form, *Maclurina* (all previously referred to as *Maclurites*). Two main taxa include *M. logani* and *M. bigsbyi*. In both cases these taxa are generally found above the Millbrig K-Bentonite in Minnesota, Illinois, Iowa, Tennessee, and Virginia. *M. logani* is found in the Bobcaygeon Formation of Ontario, the Nealmont Formation of Pennsylvania, the upper Greencastle member of the Chambersburg Formation of southern Pennsylvania, and the Chaumont, and Selby Formations of New York. These rock units have generally been assigned to the uppermost Turinian to lower Rocklandian based on other fauna including *P. undatus* (Sweet, 1984) and a number of brachiopod and bryozoan taxa (Kay, 1931). Aside from the basal Trenton, macluritids as a collective group does not make a significant reappearance in the GACB until the Edenian when it is known from a few occurrences in the Upper Mississippi Valley and in the Cincinnati Arch region. The only known exception appears to be that from the middle Trenton Farr Formation of the Timiskaming graben area in Ontario. These forms are referred to as *M. manitobensis* and are most similar to forms from west of the continental arch.

Additional taxa used in this analysis included a number of species of the strophomenid brachiopod, *Sowerbyella* as well as the orthid, *Hesperorthis tricernaria* that is now known to be present in at least four recurrence intervals from the earliest Turinian through the Edenian. These brachiopods are commonly associated with fine to coarse-grained, wackestone to packstone facies when they occur.

Other key taxa in the Chazy - Black River groups include a number of early coral and stromatoporoid genera. The former include several early types of tabulate corals including the first *Tetradium* corals: *T. cellulosum, T. fibratum, T. syringoporides, T. clarki*, and *T. columnare*. Although these taxa are widespread in middle Chazy and Black River intervals across much of the eastern GACB, they are commonly restricted in their facies of occurrence. They rarely occur with high-diversity assemblages instead they tend to be found in shallow carbonate micrite facies that are often mud-cracked and peloidal. In some instances, they have been found to be associated with *Girvanella* nodules (filamentous cyanobacteria), LLH and SH-style stromatolites, and low diversity assemblages including small bivalves, small twig bryozoans, as well as the ostracods *Leperditia fabulites* and *Eoleperditia*.

Contrary to the restricted *Tetradium* taxa, a number of coral forms including the tabulate *Foerstephyllum halli,* the rugose corals *Streptelasma corniculum, and Lambeophyllum profundum (*synonym*: S. profundum)*, as well as a labechiid stromatoporoids *Cryptophragmus* *antiquatus*, and *Stromatocerium rugosum* are characteristic of slightly deeper water, well circulated environments with higher diversity assemblages. These latter taxa are characteristic of many of the recurrent incursion faunas recognized and identified in **figure 4**.

Trilobite Biostratigraphy

Trilobite genera are also important, and include *Bathyurus*, *Cryptolithus (C. tesselatus, and C. bellulus)* and *Triarthrus*. The cryptolithid taxa are differentiated on the basis of the number of pits (fenestrae) on the glabella. Lespérance and Bertrand (1976) recognized Shermanian cryptolithids (*C. tesselatus*) were restricted in the number of glabellular pits (fewer numbers of pits). In contrast, Edenian-aged forms including *C. bellulus* were much more variable in the number of glabellular pits with several morphotypes recognized on the basis of the number and arrangement of these fenestrae. In general, all morphotypes from these younger taxa had greater numbers of them. In addition to their higher stratigraphic position, most *C. bellulus* are found in coarser sedimentary rocks (calcisiltites) as opposed to calcilutites (*C. tesselatus*). Thus the taxa in early to mid-Shermanian strata (*C. americanus – O. ruedemanni* zones) are likely all *C. tesselatus,* as is reported, while younger occurrences (*C. spiniferus - G. pygmaeus*) are likely *C. bellulus,* including those found in the Kope Formation of the Cincinnati Arch region, as well as the Reedsville Formation of central Pennsylvania.

Although excellent for recognizing time equivalency in deeper water facies in the Taconic foreland basin of Quebec, New York, Pennsylvania, central – northern Virginia, and in the Sebree Trough adjacent to the Cincinnati Arch region – the lack of these taxa in shallower water environments around the rest of the GACB area precludes their use for more significant correlations. Nonetheless, it appears these taxa immigrated into the Taconic Foreland basin during the Rocklandian highstand likely from regions in the Northwest Territories of Canada and the Mackenzie Mountains of Alaska.

The main taxa for *Triarthrus* include *T. becki* and *T. eatoni*. In general these two species are found in distinct intervals – *T. becki* appears in middle to upper Trenton rocks in New York and Pennsylvania – (usually found in calcisiltstones and fine-grained shaly calcilutites). *T. eatoni,* in turn, is found within the Edenian-aged Indian Castle Shale, and Kope Formation, and *G. pygmaeus*-bearing Collingwood Shale of Ontario and Michigan. A couple of additional species are known above the Collingwood from the Whitby Formation. These are called *T. spinosus* and T. *canadensis.* Neither of the former taxa are known below the Trenton in eastern North America. The earliest *Triarthrus* in the Upper Ordovician of North America appears to be an indeterminate species in Chazy rocks from the Mackenzie Mountain region of Alaska and the Northwest Territories of Canada. There is a species known from the Arenig of Nevada, but it is likely that *Triarthrus* did not immigrate into the Appalachian Basin until mid-Mohawkian time. The appearance of *Triarthrus* in early to middle Shermanian (upper Denley, Flat Creek Shale of New York, and Bushkill Shale Member of the Martinsburg Formation in Pennsylvania) appears to be a synchronous event and likely reflects the second opportunity for migration of these deep-water species across the transcontinental arch, or through other changes in oceanic circulation that enabled dispersal of these forms.

As for trilobites useful for correlation in the Black River Group, the characteristic genus is *Bathyurus*, a proetid. There are generally four recognized taxa. These include *B. extans, B. superbus, B. johnstoni*, and *B. trispinosis*. *B. extans* is found in the Chazy of the Mackenzie Mountains of western Canada and also appears with faunas of the Chazy Group of New York as well as the uppermost Burkes Garden Formation (Gratton Limestone sub-member) of the Stones

River Group in Tennessee. It is not known if *B. extans* emigrated from the west or if it evolved from *B. angelini* in the Champlain – St. Lawrence area and expanded to the west. It is possible that *B. extans* evolved in eastern North America from *B. angelini* as it is known from the upper Beekmantown Group (Whiterockian) in both Newfoundland and the Ottawa River region of Quebec.

Nonetheless, *B. extans* recurs, and is the most predominant bathyurid in the Black River Group. It has been collected from the Upper Carters Formation of Tennessee, the Tyrone Formation of Kentucky, the Gull River of Ontario, the Linden Hall Formation (Curtin Limestone& Oak Hall Limestone) of Pennsylvania, the Chambersburg Limestone of southern Pennsylvania, and the Lowville Formation of New York. Sister taxa, including *B. superbus* and *B. johnstoni,* are known from assemblages containing *B. extans* in the Upper Gull River of Ontario. One additional occurrence of *B. superbus* is reported from the upper Bromide (Corbin Ranch) of Oklahoma which, according to Sweet's (1984) conodont biostratigraphy, is equivalent to the Tyrone of Kentucky, and the Lowville of New York. Thus, it appears that *B. extans, B. johnstoni*, and *B. superbus* are restricted to Turinian-aged strata. The fourth taxon, *B. trispinosis*, is the uppermost taxon found. It is found only in a few outcrops in the lower to middle Bobcaygeon Formation of Ontario. It is found associated with *M. logani, H. tricernaria, P. simulatrix* and other basal Trenton forms. This is the last occurrence of the bathyurid group and the clade goes extinct in the mid Rocklandian (Sloan, 1991).

<u>Ostracod Biostratigraphy</u>

Another arthropod group that should be considered, at least briefly includes the ostracods. Most biostratigraphers would agree that ostracods are inherently very difficult to use in biostratigraphy. Like many other benthic taxa, their appearance in the fossil record is often facies dependent and in some cases they are highly endemic – especially considering their association with restricted facies. Moreover they are problematic because their preservation is often very poor, their morphologies are often very difficult to observe in the field, and there are few specialists available to confer with regarding the identification of key forms. Nonetheless, they have been studied in the past, especially the ostracods of the Chazy-Black River interval, with several key results for regional correlation of this interval (Swain, 1957; 1962).

In his two volume series on the "Early Middle Ordovician Ostracoda of the Eastern United States," Swain recognized and described dozens of taxa from the type Mohawkian region, the Appalachians, and into the Cincinnati Arch region. As mentioned, many of these taxa were newly recognized and endemic to limited outcrop regions. This appears to have been especially true during the mid to late Turinian (upper Black River Group) but also was true for some taxa during the Ashbyan interval. However, Swain (1962) did recognize two important ostracod faunal zones that he used for more regional correlation. The lowest of his faunal zones was referred to as the *Bullatella kaufmanensis* Zone and the upper faunal zone was referred to as the *Monoceratella teres* Zone. These zones were constructed on the co-occurrence ranges of several different conodont taxa that were constrained predominantly to these particular zones. That is their appearance was restricted to roughly these intervals and did not occur above or below their assigned zones with minor exceptions. Other ostracod taxa were through ranging, and did not appear useful in biostratigraphic studies.

A key observation from this study was that few ostracod taxa were holdovers from the pre-Knox unconformity Beekmantown Group (Swain, 1962). Moreover, coarse-grained siliciclastic-dominated strata immediately above the Knox Unconformity hold very few ostracod taxa. The simultaneous first appearance of a large number of ostracods occurs in the upper Day

Point Limestone of New York and demarcates the base of the *Bullatella kaufmanensis* Zone and range upward into the Crown Point Formation. These same taxa are known from the Loysburg Formation (Milroy and Clover Members) of central Pennsylvania and from the Row Park and New Market Limestones (St. Paul Group) of the Cumberland Valley of south-central Pennsylvania and Maryland. The second zone, the *Monoceratella teres* Zone, initiates in the uppermost New Market Limestone of Pennsylvania and ranges up into the Shippensburg Member of the Chambersburg Formation. Ostracods from this zone are also reported from the Hatter Formation (Eyer, Grazier members) and from the Valcour of New York. In addition, many of the taxa that make up this latter ostracod zone are also found in the basal Edinburg Limestone of Virginia.

Swain (1957; 1962) noted that ostracod biofacies from Pennsylvania, Virginia, and Oklahoma were most similar to each other. Likewise, he noted that type-Chazy forms from the Lake Champlain region were more closely allied with Baltic forms than they were to associations in the former region. This observation is interesting due to the relative geographic proximity of the Appalachian sections to the type Chazy – it would appear that some physical or environmental barrier limited the exchange and allowed some isolation between these fairly closely-spaced regions. Nonetheless, there are a few taxa that do occur in each region suggesting that the barrier (whether topographic, climatic, etc.) was only intermittently active. Ostracod assemblages from central Pennsylvania (Loysburg Formation) show an interesting and dramatic incursion event at the base of the Clover Member coincident with a deepening event. Even though subjacent Milroy lithofacies show significant similarities with the time-equivalent, higher-diversity, Row Park, ostracod diversity is relatively depauperate in the Loysburg until the flooding event. Thereafter, ostracod biofacies are more diverse; nonetheless Swain's Oklahoma-

Pennsylvania associations are still more similar than Pennsylvania – New York associations suggesting perhaps another important barrier was active.

Echinoderm Biostratigraphy

Other important contributors to the recurrent faunas recognized in the Chazy-Black River-Trenton interval are the diverse echinoderm assemblages that are widely known among collectors and paleontologists. Sprinkle and Guensburg (1997) indicate that echinoderm disparity reached an all-time high in the Turinian to Rocklandian interval in terms of class-level diversity with up to fourteen classes occurring together in some assemblages. This interval also saw the first peak in echinoderm generic diversity with substantial reduction in echinoderm diversity at the end of the Turinian. Throughout the CBRT interval, there are several key intervals that are recognized to have substantial diversity of many groups of these echinoderms. In general there are six major assemblages recognized that include: Chazy-Crown Point, Benbolt – Burkes Garden, Lebanon Limestone, Platteville Limestone, Curdsville Limestone, and the well-known Kirkfield (Upper Bobcaygeon). In most cases, these high diversity assemblages coincide with the recurrent faunas mentioned previously.

As a representative form, the diploporitan *Echinosphaerites aurantium* is represented on **figure 4** although a number of forms might also be included. *E. aurantium* is especially widespread across North America and the Baltic region. These forms are recognized either in the abundance of disarticulated plates or as entire theca that apparently resisted at least minor disturbance (Parsley and Prokop, 1986). *E. aurantium* of North America is first known from occurrences in the uppermost Benbolt Formation (Burkes Garden Member) of Virginia where it occurs with the diploporitan *Eumorphocystis multiporata*. Both of these forms also co-occur in the Mountain Lake Member of the Bromide Formation of Oklahoma. *Eumorphocystis* does not again show up in faunal lists until an unspecified species of this genus reappears in the Edenian-aged Wise Lake Formation of Iowa and Minnesota (Frest et al., 1999). Nonetheless, *E. aurantium* is found to persist in eastern Laurentia where it is found in abundance in the vicinity of the Sevier Basin. An unspecified form is found in the Wardell Limestone of Virginia and Tennessee, the Liberty Hall facies of the Edinburg Formation of Virginia, and in the Chambersburg Group (lower Shippensburg Member). *E. aurantium* or the indeterminate form does not appear as part of the Lebanon Limestone (Guensburg, 1982; 1984) recurrence interval nor does it appear in the Platteville associations described by Kolata (1975).

E. aurantium makes a return appearance in the basal Rocklandian where it is listed in faunal lists for the Oranda Formation (of the Edinburg Group) of Virginia, the uppermost Greencastle Member of the Chambersburg Group, and the Centre Hall and Rodman Members of the Nealmont Formation (Kay, 1944). *E. aurantium* is also recorded above the Millbrig K-bentonite from the Kimmswick Limestone of Missouri through Iowa and potentially into Minnesota. It occurs in the Dunleith Formation (using the terminology of Willman et al., 1975; Nelson et al., 1996) which is equivalent to the Ion member (Buckhorn-St. James beds) of the Galena Group. This echinoderm is not recorded in the Curdsville of Kentucky nor the Kirkfield of Ontario.

Another important echinoderm group, and interestingly not usually associated with *Echinosphaerites* assemblages, are the diplobathrid crinoids from the genus *Cleiocrinus*. Webster (2002) recognizes up to twelve different species of this genus within the CBRT of eastern North America. Most of these taxa tend to be restricted and may represent locally endemic populations that persisted within a small region for a limited time. Nonetheless members of this genus reappear as parts of recurrent faunas during periods of rapid sea-level rise and are usually associated with coarse-grained packstone to grainstone facies that show evidence of significant wave agitation.

There are a few taxa that show some overlap between localities. *C. regius* is known (along with two other forms) from the Mountain Lake Member of the Bromide Formation of Oklahoma and is very similar to a form (*C. perforatus*) found in the Valcour Limestone of the Chazy Group of New York. There are several forms known from Black River equivalents, but *Cleiocrinus tesselatus* is known from the Lebanon Limestone of Tennessee, the Wardell Limestone of Virginia on the Tazewell Arch, and from the Platteville Limestone of the Mississippi Valley. *C. regius* is again reported in type Kirkfieldian to Shermanian-aged strata in the upper Bobcaygeon and Verulam Formations of Ontario, the Hull Limestone of the Ottawa Valley, and from the Dunleith Formation (Rivoli Member) of Iowa. Up to seven additional taxa are described – most of which occur in the Lebanon Limestone of Tennessee (three taxa) and the Cobourg (Lindsay) formation of the Ottawa region (3 taxa). Another form (*C. sculptus*) is recognized from several intervals in Kentucky beginning in the Curdsville Limestone and extending upward into medial Lexington Group strata. It is clear that as a group – these taxa show some degree of diachroneity and therefore cannot be used alone to establish regional correlations, but collectively their recognition as part of diverse echinoderm communities tied to specific facies changes helps to constrain additional recurrent intervals.

<u>Sponge Biostratigraphy</u>

Another taxon perhaps worth mentioning – and not pictured on figure 4 is the hexactinellid sponge *Brachiospongia*. Although the occurrence of these taxa is still poorly documented, there are four known members of this genus from Rocklandian to Shermanian-aged strata. Additional taxa are known from the Edenian but the oldest known occurrences (worldwide) are the following taxa: *B. digitata*, *B. hullensis*, *B. tuberculata*, and *B. bifurcata*. The former taxon (*B. digitata*) with 9-10 radiating arms has been documented from the Brannon Member of the Lexington Group as well as from the basal Cobourg of the Ottawa Valley of Ontario where it was recognized by Wilson (1949). Wilson also described additional species: *B. bifurcata*, and *B. hullensis* from the underlying Hull Limestone. Subsequently, Rigby (1970) recognized an additional form (*B. tuberculata*) with 7-8 radiating arms from the lower to middle Bobcaygeon of the Kirkfield region of Ontario. This particular form is most similar to forms found recently in the Curdsville Limestone of Kentucky, and known for some time from the Hermitage of Tennessee (Wilson, 1949). The appearance of these taxa helps provide additional correlation tie points between Ontario and central Kentucky where at least a portion of Bobcaygeon is thought to be equivalent of the Curdsville, and the Brannon of Kentucky maybe equivalent to basal Cobourg of the Ottawa Valley.

<u>Bryozoan Biostratigraphy</u>

A final group to consider is the Bryozoa. Much work on this topic has been done by a number of workers including Ross (1963a, b, 1967a, b, 1970). There are at least two important taxa that have proven useful in more regional correlations. First, a group of distinctive cystoporate bryozoans called *Constellaria* – very well known in Edenian-aged rocks is also known from rocks from the Chazy, lower-middle Black River, and the middle-upper Trenton. Ross (1963 a, b) identified *C. islensis* (*Stellipora* of older literature is a synonym) from Isle La Motte, Vermont. This form, common in the Valcour Member of the Chazy, is thought to be very similar to forms (C. sp. & *C. lamellosa*) found in Tennessee in the Pierce and Ridley Formations of the Stone's River Group and in the Wardell Limestone (Bed 7) of Virginia.

This genus is not again listed in faunal records until the *C. americanus - O. ruedemanni* zones, where it appears in the Dunleith Group of Iowa (medial Galena equivalent), the Brannon and Stamping Ground cycles of the Lexington Group of Kentucky (Coates et al., 2004), and the uppermost Bigby-lower Cathey's Shales of the Nashville Dome of Tennessee (Wilson, 1949). *Stellipora* is recorded from an interval in the Trenton in Pennsylvania (but not specified) (Butts et al., 1939; Butts, 1945) and from the medial Trenton of the Lowville region of New York. In fact, Hall's (1847) illustrated specimen was collected from what is likely the Denley Formation from that region. These occurrences thus help to establish again some important correlations for the study region such that the earliest occurrence in the upper Chazy helps strengthen the assessments and correlations of previous authors including Cooper and Cooper (1946) on the basis of brachiopod species as does the return occurrence of *Constellaria* in the medial Trenton and its equivalents across the GACB.

The last taxon to be considered here is that of the *Prasopora.* While the specific details and morphologies of individual taxa within this group are still in need of significant work, it is readily apparent that *Prasopora* is one of the most easily recognized of all bryozoan taxa from the Ordovician. Its size and shape enables it to be found in most outcrop settings, even where heavily tectonized as is the case in the uppermost Jacksonburg Limestone in the vicinity of the Greenwich slice of the Hamburg Klippe. As noted by Sanders and colleagues (2002) *Prasopora* is usually facies specific – that is it is usually found in high abundance assemblages in mixed siliciclastic mud-rich intervals and tends to have significantly lower abundances in clean carbonate lithologies. On this basis, several recurrent intervals or epiboles, well-known from the upper Mississippi Valley, have been studied and used for correlation in that region for some time (Sanders et al., 2002). Furthermore, time-equivalent strata from Michigan, New York, Ontario, Quebec, Pennsylvania, Kentucky, and Tennessee also contain intervals with abundant *Prasopora*. Ross (1966; 1967, 1970) spent some time evaluating the occurrence of these trepostome bryozoans and their stratigraphic and evolutionary relationships in the Trenton.

At present, there are at least eighteen different species of *Prasopora* defined for the Upper Ordovician of eastern North America. Cuffey (1997) and Arens and Cuffey (1989a, b) indicate that most individual colonies of *Prasopora* actually possess a range of morphotypes and therefore support the synonymy of many of these taxa, but they indicate that there are clearly some differences that warrant at least several different form species. Nonetheless, the majority of these species are Rocklandian through Shermanian in age with additional species (not considered here) from Edenian-aged strata.

The earliest occurrence of *Prasopora* (*P. fritzae*) is reported from deep water shales of the Bromide Formation (Simpson Group) of Oklahoma (both in the Mountain Lake and Pooleville Members) by Fay and Graffham (1982). Although Cuffey and Arens are dubious of these occurrences based on conceived correlation issues, the oldest reported occurrence of *P. simulatrix* is from below the Deicke K-bentonite in the Platteville Limestone and Quimby's Mill Formation (Karklins, 1987). Based on biostratigraphic and lithostratigraphic correlation, the basal Mountain Lake Member of the Bromide is a quartz-rich equivalent of the St. Peter Sandstone, while the upper Mountain Lake is the equivalent of the Glenwood of Missouri, and Iowa. The Pooleville Member of the Bromide (which contains *E. quadridactylus* conodonts) is an equivalent of the Platteville-Quimby's Mill Member. Thus the occurrence of *Prasopora simulatrix* in the Platteville is likely contemporaneous with the occurrence of *Prasopora* sp. in the Pooleville Member of the Bromide Formation. Also, inasmuch as there are *P. fritzae* recorded from the lower member of the Bromide (in the Mountain Lake Shales) it is likely that this recorded occurrence is in fact the earliest occurrence of this entire group.

Additional occurrences of *Prasopora simulatrix*, although not recorded in the detailed study of Sanders and colleagues (2002) on *Prasopora*, are also recorded in the Upper Mississippi Valley region above the position of the Deicke K-bentonite and within the upper Spechts Ferry Shale (Glencoe Member) surrounding the Millbrig K-bentonite (Karklins, 1987). The Spechts Ferry is a bryozoan-rich environment and appears to have been significantly influenced by local to regional topographic changes (subsidence) in the Illinois Basin at this time. This form is also recorded from a few specimens in the Guttenberg and a few are recorded from the Ion Member (Buckhorn – St. James sub-members) of the Dunleith Formation (Sanders et al., 2002). In this region, however, the main occurrence or epibole, also referred to as the "*Prasopora* Zonule" (Kay, 1929) is found below the position of the hardground contact between the St. James sub-member of the Ion and the Beecher Member of the Dunleith. Superjacent to this main occurrence, two additional recurrent zones are also noted from the Eagle Point and Sherwood members of the Dunleith. In all cases, most of these are reported as *P. simulatrix* although *P. simplex* is an additional form recognized in the Decorah (Guttenberg – Ion interval).

The first occurrence of *Prasopora* in the type Black River – Trenton region is in the Selby, Napanee, Middle Bobcaygeon, and Rockland (Rocklandian strata of Ontario, and Quebec) (Liberty, 1969, Titus & Cameron, 1976; Ross, 1970). These are clearly dominated by *P. simulatrix* as well as a couple of local taxa in Ontario (*P. grandis*, *P. insularis*). This first occurrence, synchronous with the first major pulse of siliciclastics into the region during the Vermontian phase of the Taconic Orogeny, demarcates the first epibole interval for these taxa in the type Trenton region. Although *Prasopora* are found in the superjacent Kirkfield (Upper

Bobcaygeon)/Kings Falls/Hull, their abundance is significantly diminished in fossil assemblages. The second and the most pronounced epibole is demonstrated in the Sugar River Formation of New York (lowest Shermanian) and in the basal Verulam of Ontario where significant numbers of Prasopora are found associated with hardgrounds and interbedded shales in the Gamebridge Quarry of the Lake Simcoe District. This incursion of *P. simulatrix* is also recognized in the Glens Falls Limestone (Shoreham member) of the southern and central Lake Champlain region (Ross, 1970). This event is recognized in cores from Michigan (Sparling, 1964) and is likely the equivalent of the main *Prasopora* Zonule of the upper Mississippi Valley so well studied by Kay, (1929) and more recently by Sanders and colleagues (2002).

Outside of the these regions, *Prasopora simulatrix* is recorded in faunal lists from the Salona, Coburn, and Jacksonburg Limestones of Pennsylvania, the Martinsburg of Virginia, the Logana of Kentucky, and the upper-Curdsville to Hermitage of Tennessee. Moreover the occurrence of *P. falesi* in the Macedonia (Grier), Brannon, Sulphur Well, and Tanglewood of the Lexington, and *P. contigua* in the Bigby of the Nashville Dome are now considered to also represent *P. simulatrix* as per studies by Marintsch (1981) and Cuffey and Arens (1987).

Additional occurrences of *Prasopora* are recorded from uppermost Trenton units where Ross (1967b, 1970) has indicated that *P. simulatrix* underwent species splitting events into at least three additional species (*P. selwynii, P. sardesoni,* and *P. shawi*) and perhaps as many as four or more additional taxa outside of New York. Noting the commentary by Cuffey and Arens (1987), several of these forms could represent intraspecies differences, although these authors support at the least the work of Ross (1967b; 1970). In fact, on the basis that most (99%) of gum drop-shaped bryozoans from the Coburn of Pennsylvania appear to be *P. simulatrix*, they feel that the Coburn is entirely older than the upper Trenton intervals of New York and Ontario that contain the diversified forms. Nonetheless, the diversification event documented by Ross is recognized as post early Shermanian (Cobourgian in older literature) and is therefore time equivalent to the upper Verulam to lower Lindsay of Ontario, Cobourg of the Ottawa Valley, and the Rust Formation of the type Trenton region. *P. shawi* first occurs in the basal Russia, *P. selwynii* first shows up in the uppermost Russia – basal Rust, while *P. sardesoni* first appears in the middle of the Rust and is the main form found in the Steuben equivalent in the Dolgeville.

These same forms are recognized by Sparling (1964) from cores in the upper Trenton of the Michigan Basin, and *P. selwynii* is thought to be the form found in the Sherwood Member of the Dunleith Formation. Other forms including *P. patera* are found in the Lindsay Formation of Ontario and *P. nodosa* is recognized from the Catheys of Tennessee and the Lexington of Kentucky. It remains to be seen if these forms are additional endemic forms produced during the diversification event or if they are synonymous with one or another of the previously mentioned taxa. *P. simulatrix* itself is thought to persist through the speciation event and is noted from Edenian-aged strata in the Cincinnati region – yet it is not clear where the other taxa from this lineage continue.

Thus, based on correlations into the type Black River-Trenton region, it appears that the first major widespread epibole horizon occurs in the Rocklandian (Selby-Napanee, Middle Bobcaygeon, Salona, Logana, Hermitage, Buckhorn) and the second and largest epibole event, the "Prasopora Zonule", occurs in the earliest Shermanian (Sugar River, Lower Verulam, Middle Martinsburg, Grier, Beecher, Upper Hermitage). Subsequent epibole occurrences are less distinct but coincide with the diversification events of the late Shermanian. In each case, the high-abundance of *Prasopora* is coincident in many cases with hardgrounds and with the occurrence of increased shale sedimentation. Thus the initial appearance, or at the least the acme zone of each occurrence, appears to coincide with major events that are recorded across wide portions of the GACB. From outcrop analyses, these appear to coincide with significant periods of sea-level rise recorded initially by sediment starvation and hardground development followed by mud burial events that entomb many of these bryozoan occurrences. Moreover, studies by Sanders and colleagues (2002) and Cuffey and Arens (1987) suggest that disturbance may be an important component in the paleoecology of these assemblages. In addition the appearance of these taxa in strata east of the Illinois Basin and upper Mississippi Valley, both on the GACB platform areas and in the Taconic Foreland, is likely due to the submergence of much of the GACB and the development of somewhat deeper water connections.

EVENT STRATIGRAPHY

In addition to biostratigraphic correlations discussed previously, there are a number of additional stratigraphic horizons and intervals that are useful in establishing a highly resolved chronostratigraphic framework. The following discussion identifies and integrates a number of important event horizons from the Chazy, Black River and Trenton intervals from across eastern North America. Correlation of these horizons is constrained initially by the graptolite and conodont biostratigraphic framework discussed previously. Secondarily, it is constrained using additional short-term, high-frequency events that are recognized over local to regional areas. These short-term, high-frequency events take the form of important sequence stratigraphic surfaces (including sequence boundaries, facies dislocation surfaces, maximum flooding surfaces, etc.), chemostratigraphic events (neodymium, and carbon isotopic excursions), lithostratigraphic changes, and other unique markers discussed elsewhere.

K-bentonites

A significant and important method for correlation in CBRT strata have been the large number of altered volcanic ash deposits that have been known from these units and studied for nearly a century. Meta-bentonites or potassium bentonites (K-bentonites) have been studied extensively on the basis of their: 1) bio-stratigraphic occurrence, 2) mineralogical composition, and in some cases 3) their chemical fingerprints (Delano et al., 1990; Haynes, 1994; Kolata et al., 1996; Mitchell et al. 2004; Carey, 2006) and correlated across eastern North America. Recent and historic studies, (Kay, 1935; Adhya, 2000; Cornell, 2001) have demonstrated the presence of K-bentonites that are potential correlatives of those correlated widely by Haynes (1994), Kolata et al. (1996) and others. Moreover, these horizons have been identified in subsurface wire-line logs across the Mississippi River Valley, Cincinnati Arch and into West Virginia (Kolata et al. 1996, Huff & Kolata, 1990) (Figure 13). At least one of them is now reported in the type Black River region (Mitchell et al., 2004). **Figure 5** shows the relative distribution of K-bentonites in the major outcrop regions of this study. K-bentonite positions are constrained using known chemically finger-printed K-bentonites, available biostratigraphic control, and position relative to other key marker positions as established herein. In the New York – Ontario type region, no less than 24 K-bentonites have been recognized, and in central Pennsylvania Kolata et al. (1996) report no less than 29 K-bentonites for the same interval.

This stratigraphic study relies primarily on the identification of K-bentonites as established by previous authors for individual outcrop areas. The contribution of this study to K-bentonite event stratigraphy lies in three primary capacities: 1) establishing the contextual lithostratigraphic positions of K-bentonite correlations between outcrop regions evaluated in this study (some established previously by Conkin, 1991, and the large regional synthesis of Kolata

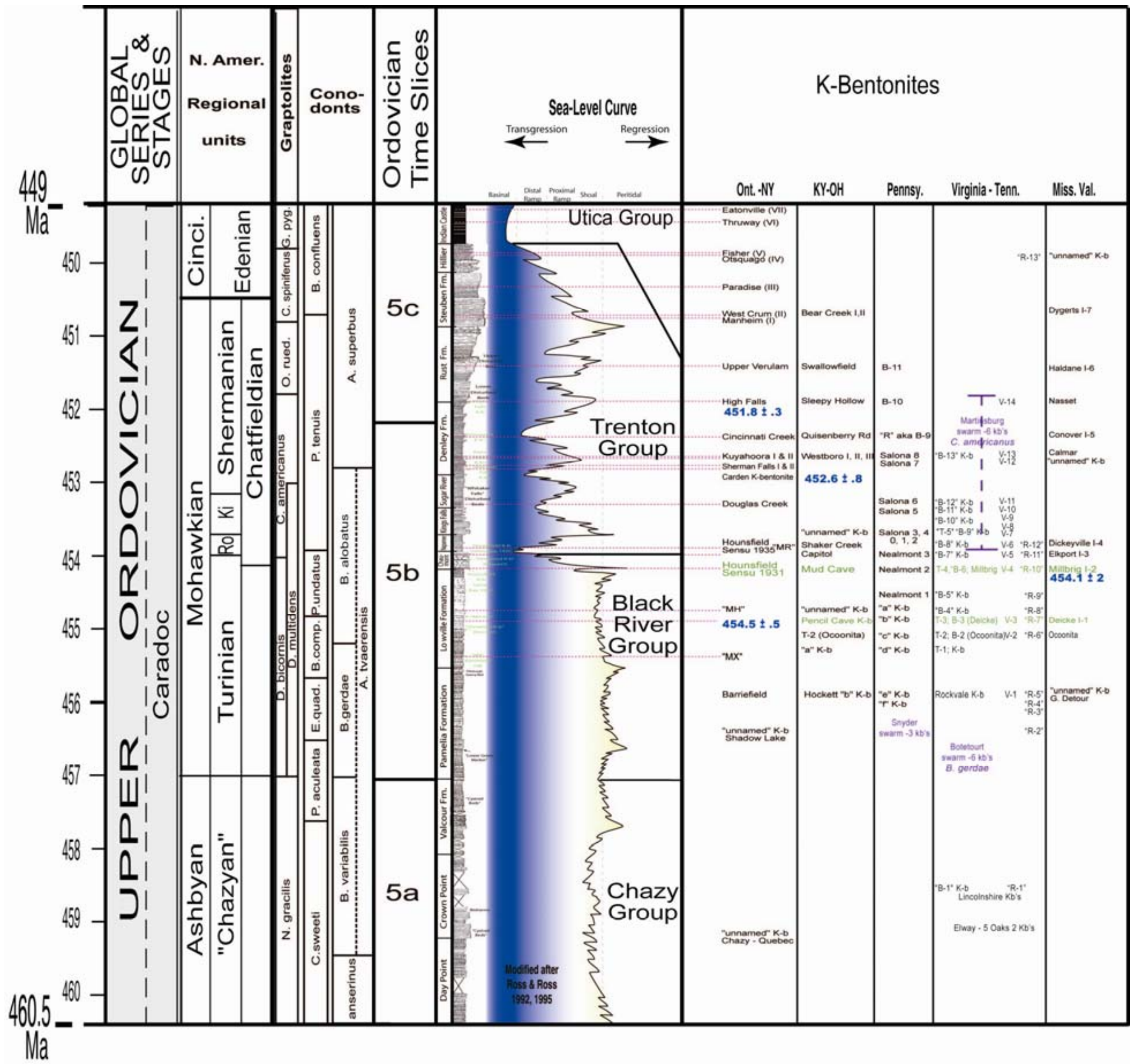


Figure 5: Relative Distribution of named and unnamed K-bentonites in the Chazy-Black River-Trenton interval from Eastern Laurentia. Those K-bentonites shown in green have been correlated confidently on the basis of their geochemical fingerprint. Also shown in blue are U/Pb geochronologic ages of K-bentonites as established elsewhere.

et al., 1996) and proposing likely lateral equivalents beyond the well-known Deicke and Millbrig; 2) establishing a new reference stratigraphic section for the position of the Hounsfield K-bentonite in the upper Black River of New York State – with special attention to stratigraphic nomenclature and the recognition of two Hounsfield K-bentonites (sensu Kay, 1931; and sensu Kay, 1935); and 3) the regional significance of K-bentonite activity in the evolution of the Taconic Orogeny. The following discussion will focus on points 1 and 2 above with the 3rd point for discussion in a subsequent chapter.

As shown in **figure 5**, a large number of altered volcanic ash beds have been recognized on the basis of their field characteristics (as documented by Haynes, 1994; Kolata et al., 1996), phenocryst composition, their clay mineralogy, and the phenocryst chemistry. Among the most important K-bentonites in the literature have been the widespread and voluminous Deicke and Millbrig – both of which have been the primary focus of most recent volcanic ash studies (Emerson et al., 2004; Carey, 2006). Recognition of these K-bentonites has been tantamount for linking the GACB region with the foreland basin succession in the Appalachians (see Haynes, 1994; Kolata et al., 1996, 1998). The position of these K-bentonites is well-established and supported for correlations between the Upper Mississippi Valley, the Cincinnati Arch/Nashville Dome region, and the southern and central Appalachians.

Farther north, in north-central Virginia through central Pennsylvania the position of the Deicke and the Millbrig have been estimated by chemical studies of McVey (1993) as previously discussed. As a result, in these regions, the Deicke and Millbrig are reported from the Upper Oranda and basal Salona respectively. Although it is agreed that the K-bentonites in the upper Oranda and the basal Salona Formations are likely correlative with one another (Thompson, 1963; Cullen-Lollis & Huff, 1986), numerous lithostratigraphic and biostratigraphic observations argue against Deicke and Millbrig equivalency for these beds. In addition, although whole-rock analysis of K-bentonites was useful in local correlation of K-bentonites, Cullen-Lollis and Huff (1986) were unable to establish equivalency between central Pennsylvania and elsewhere on more regional scales. Moreover additional lines of evidence argue against these K-bentonites as representing these bentonites as well. First, the relatively close spacing of these K-bentonites runs counter to observations that show this region to have very high subsidence rates throughout much of the Turinian – the highest in the GACB region.

Second in Pennsylvania and Virginia, the position and timing of the Guttenberg Isotopic Carbon Excursion (GICE) (discussed elsewhere) does not fit the correlations of McVey (1993). All across the GACB, including in southern Virginia (Hagan) and West Virginia (Dolly Ridge), the GICE initiates after deposition of the Millbrig, not prior to it as shown in early studies (Patzkowsky et al., 1997). Young and colleagues (2005) report the full occurrence of the GICE in the base of the Trenton at Hagan and in the base of the Dolly Ridge Formation in West Virginia. In both cases, the isotopic excursion occurs above the Deicke and the Millbrig and initiates positive rise just prior to the V-7 K-bentonite and reaches maximum values just above the position of the House Springs K-bentonite (Young et al., 2005). Both of the former K-bentonites are shown to occur in the upper Eggleston Limestone and in the Nealmont Formation, respectively. Moreover recent apatite phenocryst analyses show the Hagan Deicke and Millbrig K-bentonites to again be correlative with the type Deicke and Millbrig, and the sample in the Nealmont at Dolly Ridge to be a likely equivalent of the Millbrig (Carey, 2006). Thus it appears that the K-bentonites in the basal Salona of Pennsylvania and the basal Dolly Ridge Limestone (West Virginia) occur above the position of the type Deicke and Millbrig and thus are likely miscorrelated in Central Pennsylvania. Moreover, samples from McVey's (1993) Millbrig were sampled and studied by Carey (2006), but unfortunately no apatite phenocrysts where recovered from the samples– apatites are a characteristic component of the Millbrig in nearby regions.

Thus it is proposed herein on the basis of these arguments, that the basal Salona of Pennsylvania and upper Oranda K-bentonites of the Shenandoah Valley, Virginia, should be re-evaluated in future studies. It is suggested that the basal Salona – upper Oranda K-bentonites likely represent equivalents of K-bentonites found above the Millbrig (i.e. V-5, 6, 7, Shaker

Creek – Dickeyville – House Springs – Elkport and other as yet "un-named" K-bentonites) and not depositional equivalents of type-Deicke and Millbrig. Nonetheless, some important stratigraphic progress has been made with respect to establishing the positions of these K-bentonites in the type BRT region of New York and Ontario.

K-bentonite Correlations: Deicke

In New York State and Ontario, the position of the Deicke K-bentonite is still enigmatic and not well-established, although a number of K-bentonites exist within the Lowville Formation and maybe candidates. Kolata and colleagues (1996) suggested the Deicke maybe equivalent to the MH K-bentonite of Liberty (1969) and to a K-bentonite reported by Walker (1973) from the Lowville. This K-bentonite is the thickest and most obvious of the Ontario K-bentonites. However, it is one of several K-bentonite horizons in the Gull River – Lowville succession of New York (Cornell, 2001). The MH of Ontario is characterized by a relatively high abundance of quartz phenocrysts and carries a moderate percentage of altered, euhedral biotite phenocrysts and a few other phenocryst species including zircons that have been dated (and briefly cited by Armstrong, 2000). Collectively this association is not typical for the Deicke as reported previously (Haynes, 1994; Kolata et a., 1996). The high biotite and quartz abundance are characteristic of the Millbrig and succeeding K-bentonites. Therefore the position of the Deicke in Ontario and New York is still not confidently placed.

In the subsurface outside of the New York - Ontario outcrop region, several K-bentonites have been located in the immediate vicinity of the Deicke, including in the subsurface of Ohio and in outcrops of central Kentucky (see Stith, 1979; Kolata et al., 1996, Armstrong, 2000). All of these K-bentonites are located within a relatively narrow stratigraphic interval, but most importantly where both the Millbrig and Deicke are established confidently by fingerprinting, there are one additional, persistent K-bentonite and a thinner, less persistent K-bentonite between the Deicke and the Millbrig. In fact, Stith (1979) recognized a total of five K-bentonites in the Tyrone Formation, and Conkin and Conkin (1983) suggested there were as many as eighteen in the same interval. The thickest and most persistent were referred to as the alpha, beta, gamma, delta, a, and b in descending order from the top of the formation (by Stith). Kolata and colleagues (1996) inferred these beds to represent the Millbrig, Deicke, Ocoonita, a, and Hockett K-bentonites respectively.

Nonetheless there may be an error in this latter interpretation. In wireline-logs studied by Stith, the separation between the alpha (suspect Millbrig) and the beta (suspect Deicke) is only between two to four meters. This is about half the separation measured in cores, and in outcrops of central Kentucky where the Deicke and Millbrig are placed relative to other key stratigraphic contacts. In most cores in the Cincinnati Arch region, just as in outcrops, the distance between the Millbrig and the Deicke is regularly between eight to ten meters (Ettensohn, 1992), and in almost all cases the Millbrig and Deicke are separated by an intervening K-bentonite. Thus as suggested by Huff and Kolata (1990) the alpha K-bentonite in the subsurface is likely the Millbrig, but subjacent beds should be re-evaluated given the presence of at least one if not two additional K-bentonites between it and the Deicke.

Based on outcrop studies and its characteristic stratigraphic position, the persistent K-bentonite between the Millbrig and Deicke (unnamed K-bentonite in Kentucky) is presumed to be the equivalent of the MH from Ontario (Brett et al., 2004). In outcrops of central Kentucky, the distance between the “unnamed” or MH correlative and the underlying Deicke is typically about four meters. As such, based on the close association of these K-bentonites the MH maybe used for longer distance correlations from Ontario into Ohio even though it is not the exact correlative of the Deicke bed.

In New York, recent investigations (Carey, 2006) failed to yield apatite phenocrysts from the MH equivalent horizon for chemical studies based on analysis of apatite phenocrysts. However, Carey (2006) did recognize apatite phenocrysts from multiple cores in Ohio (both southern and northern Ohio) that showed affinities for the type Deicke and Millbrig of the Upper Mississippi Valley. Moreover, in one core (OHIO DGS Core 3409), it was shown that two separate beds (~5.5 meters apart) showed apatite phenocrysts with Deicke chemical signatures. As remarked by Carey (2006), no two beds have been previously reported to have nearly identical apatite chemistries for the Deicke much less for other beds. This observation suggests the possibilities that either these beds were: 1) different eruptive events with identical chemistries or 2) were derived from the same eruptive event with subsequent remobilization and erosion of previously deposited ash materials. This particular core analyzed by Carey, was taken from the margin of the Rome Trough in southern Ohio which may have been re-activated at least locally due to late-phase Blountian tectonism and may have provided a source area for reworking of Deicke K-bentonite before deposition of the Millbrig. As reported by Haynes (1994) the Deicke appears to be missing in some localities in the southern Appalachians above the position of the Walker Mountain Sandstone. Therefore it is plausible that the Deicke could have been remobilized due to local uplift and erosion and transported into more interior regions.

Interestingly, Carey (2006) identified a possible Deicke correlative in New York State at the contact of the Watertown and Selby Formations. The assignment of this K-bentonite to the Deicke is particularly problematic due to recognition of the Millbrig K-bentonite by Mitchell and colleagues (2004) at about the same stratigraphic position in the type locality of the Hounsfield

K-bentonite (see discussion below). Again either multiple eruptive events can produce the same apatite chemistries or single event beds may be prone to erosion and re-sedimentation. As suggested by the textural appearance of phenocrysts analyzed by Carey (2006), this bed may indeed represent remobilization and erosion of a previously deposited event bed.

At Union Furnace in Central Pennsylvania, one of the basal Salona K-bentonites (Salona 2 of Thompson, 1963; B-12 of Berkheiser & Cullen-Lollis, 1986) also shows apatite chemistries plotting with the Deicke K-bentonite. However, strontium isotopic values ($^{87}Sr/^{86}Sr$) are larger than the Deicke and even slightly above the values of the Millbrig as reported from the Upper Mississippi Valley. Carey (2006) reports that the B-12 is most similar to the Deicke; however, he does not further report any textural data or SEM images of the phenocrysts analyzed in his single aliquot run of five phenocrysts. Thus as in New York, data are problematic and without additional stratigraphic support – the analysis is still inconclusive for establishing stratigraphic position of the Deicke and Millbrig in this region.

K-bentonite Correlations: Millbrig

As for the Millbrig K-bentonite in the Ontario-New York type-region, biostratigraphic evidence from Kay (1931), lithostratigraphic evidence from Conkin (1992) and chemostratigraphic evidence from Barta (et al., 2003, 2004), and from Mitchell and colleagues (2004) have supported recognition of the Millbrig as the “Hounsfield” K-bentonite in the type-outcrop region of New York State, i.e. in a position below the Watertown Limestone as designated initially by Kay (1931). The synthesis of these various data along with sequence stratigraphic assessments helps provide the correlative framework of this study – although some confusion still exists as to lithostratigraphic nomenclature and the setting of the Hounsfield K-bentonite. Again, Kay (1931) placed the Hounsfield in the Glenburnie Shale below the

Watertown Limestone, but later Kay (1935) revised the stratigraphic context of this K-bentonite and suggested that it occurred above the massive Watertown and within the Selby Limestone. In the type Hounsfield quarry there appears to be only one K-bentonite; however, discrepancy exists in terms of its stratigraphic placement, i.e. does the "Hounsfield" K-bentonite occur below the Watertown Limestone (sensu Kay, 1931), or does it occur within the base of the Selby Limestone (sensu Kay, 1935; see Brett et al., 2004 vs. Mitchell et al., 2004)? This problem is exacerbated due to three additional concerns:1) stratigraphic incompleteness of the type section quarry, 2) recognition of K-bentonites elsewhere both at the base of the Selby and below the Watertown, and 3) the position of a regional sequence boundary that truncates the base Watertown K-bentonite (Hounsfield sensu Kay, 1931) in some areas.

The Hounsfield K-bentonite from the small, poorly exposed type-section of the Hounsfield adjacent to the Game Farm Road in Brownville, New York (**figure 6**) was sampled for quartz inclusion and apatite phenocryst chemical analyses (reported by Mitchell et al., 2004). The stratigraphic context of the sampled K-bentonite was reported to occur in the base of the Selby Limestone following the assessment of Kay (1935). Unfortunately, the K-bentonite was sampled from an incomplete section without additional reference to local stratigraphic context. As reported by Kay (1931), Kay (1935), Conkin (1992), Cornell (2001), Mitchell and colleagues (2004), and herein, the exposure in the type-section containing the Hounsfield is limited to a poorly exposed ~3.5-4 meter section in the swamp-filled Farr Quarry. There is very little outcrop exposure below – except for a couple of bedding planes, and an 8-9 meter thick covered

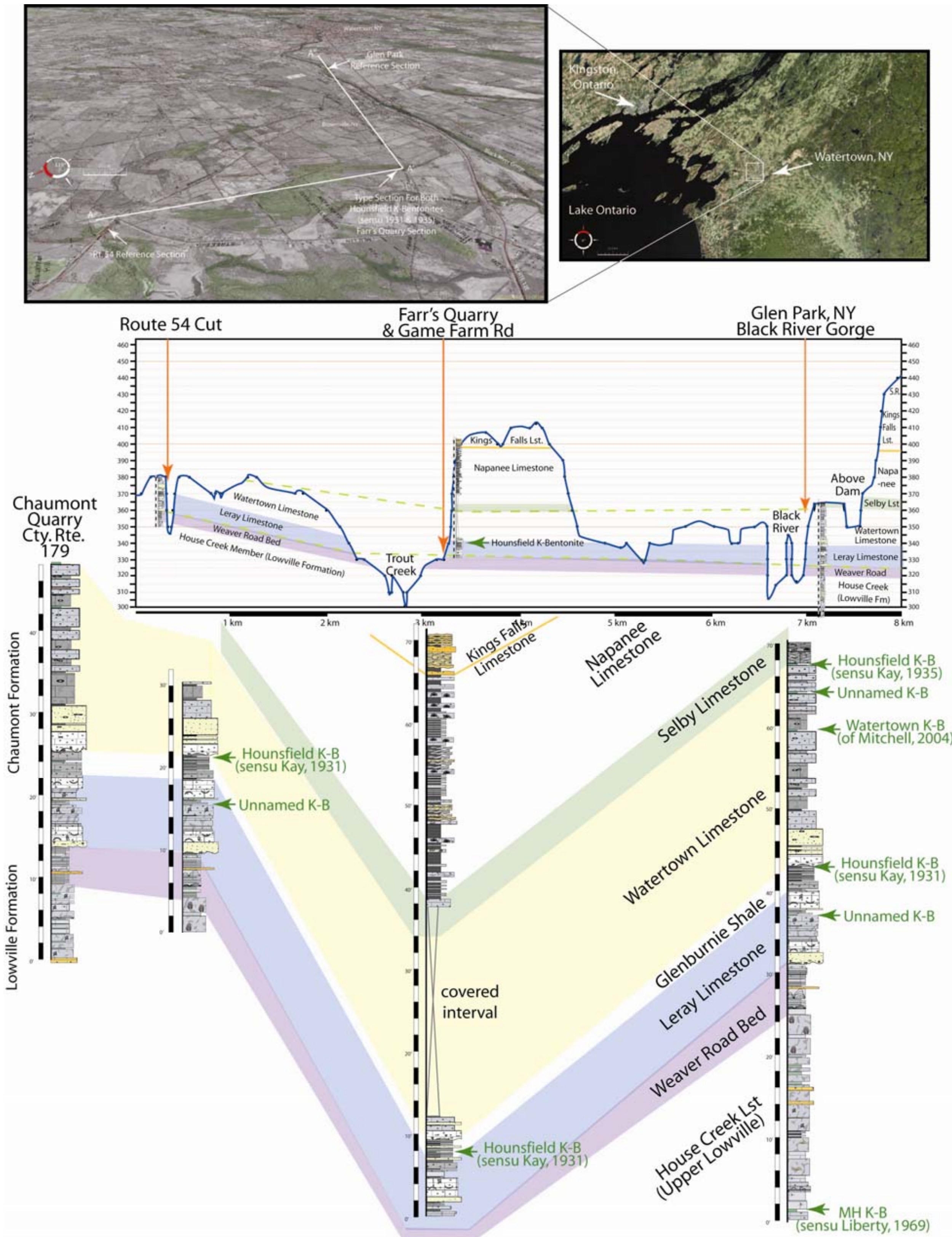


Figure 6: Stratigraphy of the Upper Black River Group in the type region of the Chaumont Formation and Watertown Limestone. Also shown is the reference section for the Hounsfield K-bentonite and representative reference sections.

Originally, Kay (1931) described the setting of this K-bentonite as occurring within the interval of his Glenburnie Shale above the Leray and below the Watertown limestones. After his original description, Kay (1935) reassigned the stratigraphic succession within which the K-bentonite bed occurred to the Selby Formation. In the latter study, the limestones below the K-bentonite were assigned to the top of the Watertown, and the beds above to the base of the Selby Formation based on faunal evidence and the recognition of K-bentonite at the base of the Selby in Ontario. If the type "Hounsfield" K-bentonite were in the Selby (and not below the Watertown) then the bed would occupy the same stratigraphic position as the basal Selby K-bentonite that was sampled at Glenn Park, New York and analyzed by Carey (2007). This sample, as mentioned above, showed a chemical signature unlike the Millbrig and instead was more similar to the Deicke. Clearly these K-bentonites are not the same, and are not in the same stratigraphic position.

Detailed mapping reported here (and shown in **figure 6**) is used to establish a more complete stratigraphic reference succession in the Watertown-Brownville region in order to fill in the covered section at Game Farm Road and to establish an accurate lithostratigraphic position for the Hounsfield K-bentonite. This stratigraphic mapping is significant because it has allowed recognition in the Watertown area of key stratigraphic intervals as originally defined in the Black River region (Cushing et al., 1910; Kay, 1931, Walker, 1973, etc.). In addition, it has allowed recognition of the complete succession of K-bentonites within the interval, showing at least five different K-bentonites that conform to field recognition criteria. This work further establishes the position of two different K-bentonites that have been reported by Kay, i.e., the basal Selby K-bentonite of Ontario, (aka "Hounsfield" sensu 1935) and the Glenburnie hosted Hounsfield K-bentonite (sensu Kay, 1931).

Proposed Reference Sections for type-Hounsfield K-bentonite

In 1999-2000 road-widening activity along county Rte. 54, approximately three kilometers north of the Hounsfield type section, exposed a ten meter-thick succession of upper Lowville through Watertown Limestone (**figure 7, lower right**). Capitalizing on the position of bed partings afforded by K-bentonites within the overall massive limestones, benches in the ditch were conveniently located at the level of most of these partings. As a result, discontinuous lenses and pockets of bentonite, including a bentonite in the position of the type Hounsfield sensu Kay 1931, were made accessible. Moreover, re-engineering work in the diversion channel for the hydro-electric plant on the Black River at Glen Park (~four kilometers east of the Hounsfield type section) required that the diversion channel be drained for a short time in 2004-2005. The vertical north-wall of the diversion channel exposed a fourteen meter succession from the Lowville through the top of the Watertown Limestone and included about one meter of the basal Selby Formation (**figure 7, upper left**). As shown in Figure 6, these exposures have helped to establish a series of easily-recognizable beds and cyclic packages that can be identified and correlated across the 7-8 kilometer area in question and beyond. These units are also recognizable in the type localities of the Chaumont Formation in the Watertown area, in the Boonville region, and in the type Leray quarry which is 4.9 kilometers east of the Glen Park locality.

Correlation between these sections is based primarily on bed-level lithostratigraphic correlation and the position of K-bentonites relative to key beds. Of particular importance has been recognition of the uppermost Lowville Formation House Creek and Weaver Road beds in

Figure 7: Outcrop photographs for four different localities in the Black River Valley, where K-bentonites are exposed in the uppermost Lowville to lower Selby Formations. Important reference localities include those exposed along Rte. 54 in the Town of Brownville, and on the Black River at Glenn Park. As shown in the diagram are the positions of at least three different K-bentonites.

the minor outcrop exposures along the length of Trout Creek and in the field northwest of the type-Hounsfield or Farr quarry. The Weaver Road is nearly two meters thick at the Brownville reference section and just slightly thicker in the Glen Park Gorge. Moreover, given a regional dip in this area of just under 10 degrees to the south west, the position of the Weaver Road-Leray contact is projected into the base of the now flooded, Farr Quarry (type-Hounsfield). Thus the assumption that the fields below the Hounsfield type quarry are actually of upper Watertown (as per Conkin, 1992) is in error, and therefore the stratigraphic assignment to Selby used by Mitchell and colleagues (2004) based on Kay, (1935) and Conkin's stratigraphic columns are also problematic with respect to stratigraphic placement.

Based on bed-level correlations with bracketing outcrops (i.e. Rte 54, and Glenn Park Gorge sections), the interval represented by the Hounsfield type-section at Farr Quarry on Game Farm Road is constrained to the Leray through basal Watertown Formation only. The sixty centimeter shaly, nodular interval within which the Hounsfield occurs is representative of the Glenburnie sections in Ontario. As in Ontario the interval is capped by a fairly prominent ledge of crinoidal-brachiopod packstone to fine-grained grainstone that grades upward into finer-grained fossiliferous wackestones that often are extremely bioturbated and cherty. This latter facies is typical of the Watertown Limestone and the upper part of the Chaumont Formation in their type localities. When correlated between the Rte 54 locality and the Glen Park gorge sections – it is clear that strata at the Rte. 54 locality are truncated before the top of the Watertown is reached and therefore no Selby is present. Likewise, only the basal Watertown is exposed in the Farr Quarry on the Game Farm Road and the upper Watertown to base Selby are covered in this location. A complete section of Watertown Limestone is afforded in Glen Park where it is capped by the Selby Limestone and the position of the contact is marked by the distinctive K-bentonite analyzed by Carey (2007) and discussed previously.

Thus based on correlations shown in **figure 6**, the position of Carey's (2006) basal-Selby K-bentonite and Mitchell and colleagues (2004) "basal-Selby K-bentonite" are not the same horizon although they conform to the modified and original descriptions for the Hounsfield.

Chemical analysis suggests that the original type Hounsfield is likely Millbrig (Mitchell et al., 2004), as claimed by Kay (1931), and supported lithostratigraphically by Conkin (1992). As noted, the superjacent basal Selby Ash of Glen Park ("Hounsfield" sensu Kay, 1935) has signatures typical of the Deicke. Given the reverse stratigraphic order (Millbrig below, Deicke above), and the occurrence of distinctly weathered apatite grains as reported by Carey (2006) in the basal Selby K-bentonite there is now plausible evidence for sedimentary reworking and re-deposition of previously deposited K-bentonites.

Post-Millbrig K-bentonites

As noted, Carey (2007) reported data from apatite phenocrysts from the basal-Selby K-bentonite at Glen Park. This bed sits in the position of the basal Selby K-bentonite ("Hounsfield" sensu Kay, 1935), and is also recognized by Cornell (2001) from outcrops in the Napanee region of Ontario where it is used to separate the Watertown Limestone below from the Selby above. The results of multiple chemical analyses of apatites indicate that this K-bentonite, which occurs some 24 meters (~80 feet) above the MH K-bentonite in the Gull River Formation, maps primarily to the Deicke K-bentonite discriminant fields on the basis of Mn vs. Mg, Fe vs. Mg (wt. %), etc. following methods used by Emerson et al., (2004) for the Upper Mississippi Valley samples. Carey (2006) reported that most apatites from the basal-Selby bed were frosted and showed, in SEM images that many had rounded morphologies suggesting that they were modified by depositional processes. Rounded samples were excluded (60 apatite phenocrysts) or were mechanically abraded (22 apatite phenocrysts) prior to analysis. One aliquot produced strontium isotopic compositions ($^{87}Sr/^{86}Sr$) similar to the Deicke. One aliquot (60 phenocrysts) had strontium isotopic compositions similar to the Millbrig, and one aliquot run of seven pristine apatites produced strontium values unlike either type Deicke or Millbrig. This thus appears to be a higher ash bed that mimics the Deicke, and or represents a much more complicated scenario than is yet proposed. This bed could correlate with the B-12 K-bentonite recognized in the basal Salona in Pennsylvania which also appears to be a Deicke mimic.

A third K-bentonite has also been fingerprinted and reported by Mitchell and colleagues (2004) as distinct from the Millbrig and the Deicke. This K-bentonite, analyzed for both apatite chemistry and quartz melt inclusions is termed the "Watertown Quarry Bed" by Mitchell et al., or "unnamed" as is labeled in **figure 7**. Mitchell does not report a specific locality for the bed, and several quarry exposures in the Watertown region have recognizable K-bentonite beds preserved within the upper portion of the Watertown Limestone, but below the position of the basal Selby K-bentonite as just discussed. The most prominent bed is well exposed three meters below Carey's K-bentonite in the Glen Park locality. It is also exposed in the type Leray quarry just north of Watertown, and in the Boonville Barrett Paving Quarry. Evidently, this K-bentonite represents one of the K-bentonites above the Millbrig and maybe the equivalent of one of the K-bentonites in the Curdsville/Guttenberg Limestone interval.

Further evidence for these correlations comes from recognition of the GICE as reported by Barta (2004) in New York. Carbon isotopic values surrounding the Hounsfield sensu stricto /Millbrig are characteristic of values reported elsewhere for the Millbrig interval. Moreover, Barta (2004) shows rapidly increasing carbon isotopic values in the Napanee Formation above the nine meter-thick covered interval at Game Farm Road (within which the Watertown-Selby interval is contained). The characteristic double peaks of the GICE are shown in the Napanee, and isotope values must climb rapidly in the intervening strata. Thus by correlation with the complete Glen Park section, the interval within which the "Watertown Quarry K-bentonite," and the "Basal Selby K-bentonite" occurs, correspond to the interval above the Millbrig. In the upper Mississippi Valley, the Cincinnati Arch, and the central Appalachian region, this interval contains the Elkport, Dickeyville; Capitol, Shaker Creek; V-7, and House Springs K-bentonites respectively. Moreover, as previously discussed the interval of the supra-Millbrig K-bentonites most effectively correlates with the K-bentonites in the uppermost Nealmont to basal Salona Formation in central Pennsylvania above the initiation-level of the GICE in that area.

CHEMOSTRATIGRAPHIC EVENTS

Recent work has focused on the identification and correlation of important chemostratigraphic horizons within the Late Ordovician – and especially within the Turinian to Chatfieldian interval (Ludvigson et al., 2000; 2004; Young et al., 2004). There are three isotopic systems that will be emphasized here (carbon, neodymium, and strontium) as they are especially important in regard to correlation of the lowermost Chatfieldian. These are discussed in detail for their implications elsewhere in this dissertation; however their stratigraphic context is described here.

Guttenberg Isotopic Carbon Excursion (GICE)

The "mid-Caradoc Carbon Isotopic Excursion" or the Guttenberg Isotopic Carbon Excursion" (GICE) is a positive isotopic excursion first recognized in the subsurface of Iowa (Ludvigson et al., 1996). Subsequently, it has been recognized in other outcrops in the same region where it has now been recognized as the most prominent of a growing list of isotopic excursions within the Turinian to Chatfieldian interval (Ludvigson & Witzke, 2005; Fanton & Holmden, 2007). Like the significant excursion in the Hirnantian Stage of the latest Ordovician, this event appears to be a global event as it has been recognized in the Baltic and in China (Ainsaar et al., 1999; Saltzman et al., 2003). In the Upper Mississippi Valley, the GICE is a slightly positive $\delta^{13}C$ excursion of 1-2‰ above pre-excursion values (typically less than 0‰). The excursion, however, becomes somewhat more amplified away from the Transcontinental Arch toward the Sebree Trough and the cratonic margin (Fanton and Holmden, 2007).

Figure 8 shows the anatomy of the GICE excursion as it is expressed in two cores from

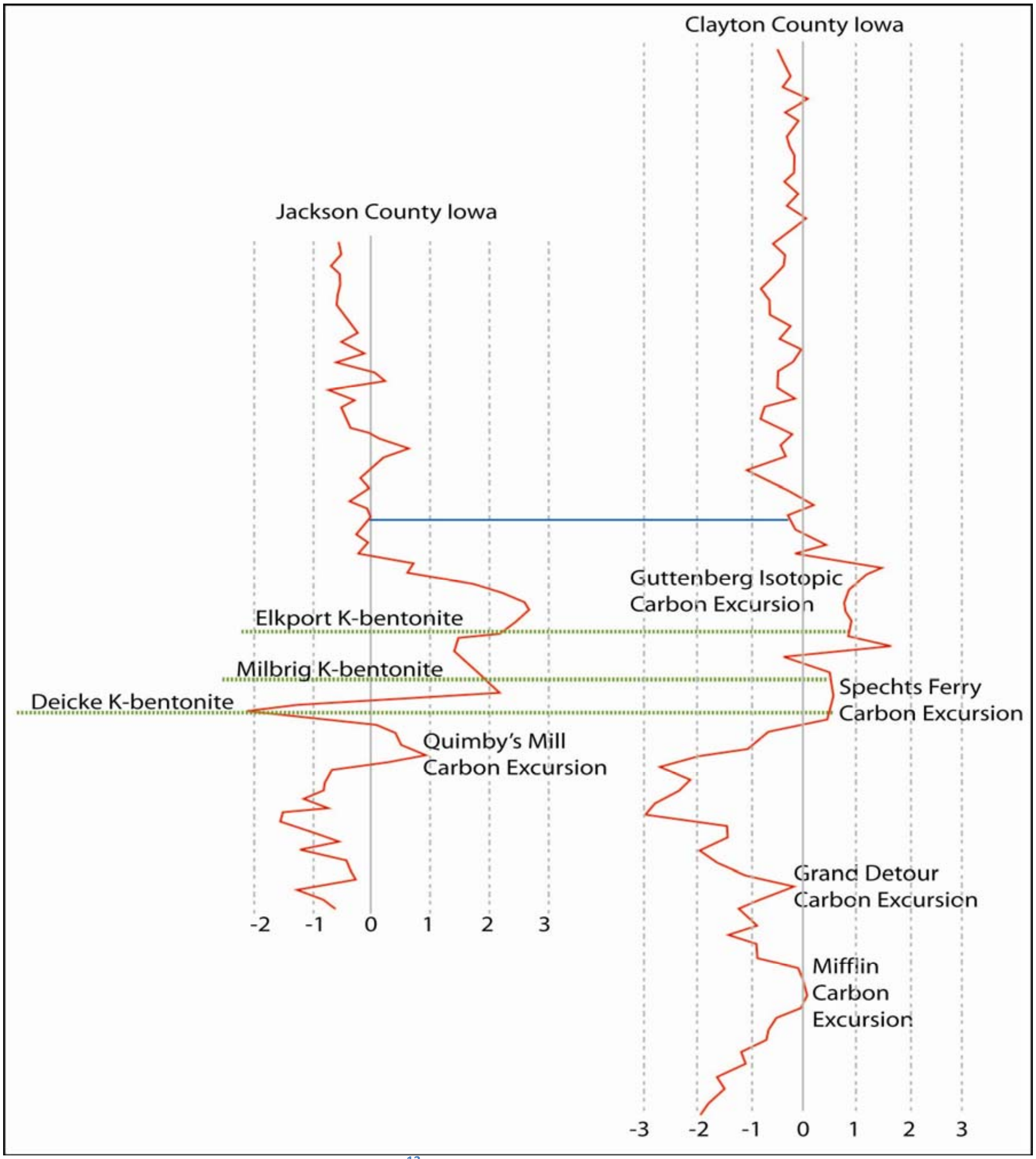


Figure 8: Carbon isotopic signatures (reported as $\delta^{13}C$) as measured from two cores in Iowa (data from Ludvigson et al., 2004). The position of the Deicke, Millbrig, and Elkport K-bentonites relative to the prominent GICE excursion is shown in addition to the position of other excursions subjacent to the GICE.

Iowa (data from Ludvigson et al., 2004). As shown, the GICE shows up as the uppermost of five carbon isotopic excursions in the transition between the Black River and Trenton Groups (Turinian to Chatfieldian). In the Clayton County core (Big Spring No. 4), the GICE is located about four meters above the Deicke K-bentonite and maximum values of the excursion are reached in the immediate vicinity of the Elkport K-bentonite. In the Jackson County core (Cominco SS-12), the excursion again occurs about four meters above the Deicke K-bentonite. In both of these cores, the Millbrig has not been recognized, although it generally occurs about two meters above the Deicke. Elsewhere in Iowa (i.e. the Elkader A1 core in Clayton County), the GICE begins its sharp increase about two meters above the Millbrig. As mentioned, the GICE reaches maximum values just above the Elkport K-bentonite and typically returns to background values near the top of the Guttenberg Member of the Decorah Formation. Stratigraphic studies suggest the total duration of the excursion to be relatively short and dominantly developed during a portion of a single depositional sequence (Witzke & Bunker, 1996; Fanton & Holmden, 2007).

Outside of the upper Mississippi Valley additional studies have outlined the anatomy of the excursion across the GACB (Fanton & Holmden, 2001, 2007; Ludvigson et al., 1996, 2000, 2002, 2004; Ludvigson and Witzke, 2005; Panchuk et al., 2005, Panchuk et al., 2006; Saltzman et al., 2001; Saltzman & Young, 2005; Young et al., 2005). The GICE is now considered to be one of ten Turinian to Chatfieldian positive isotopic excursions recognized and correlated in the Wisconsin Arch to Illinois Basin region. The occurrence of these additional excursions indicates an increasingly complex scenario for carbon cycling in the GACB (Ludvigson et al., 2004; Ludvigson and Witzke, 2005). Nonetheless, the stratigraphic position and characteristic anatomy of the GICE generally allows it to be recognized relative to the others that range from the Turinian up to the base of the Edenian stage (**figure 9**).

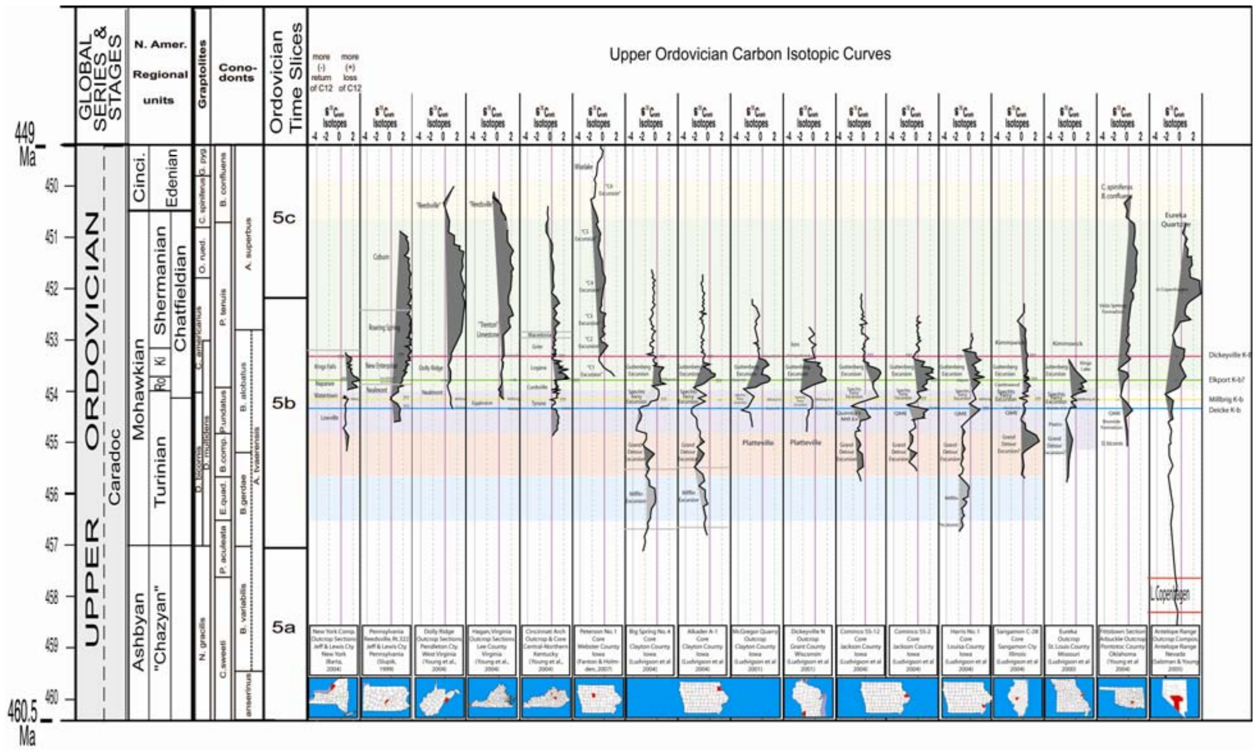


Figure 9: Carbon isotopic curves for localities across the GACB region including New York, Pennsylvania, West Virginia, Virginia, Iowa, Wisconsin, Illinois, Missouri, and Oklahoma. Also shown is the profile for Nevada. The stratigraphic position of profiles is established relative to the position of the four K-bentonites: Deicke, Millbrig, Elkport, and Dickeyville as currently established. Isotopic curves modified from data presented by: Barta (2004), Fanton & Holmden (2007), Ludvigson et al., (2000, 2001, 2004), Salzman & Young, (2005), Slupik (1999), Young et al., (2004) (see Appendix A for larger version).

Using biostratigraphy, and the position of major K-bentonites including the Millbrig K-bentonite, the GICE has now been recognized in Oklahoma, Kentucky, Virginia, West Virginia, and in New York (Young et al. 2005; Saltzman & Young, 2005). An excursion in central Pennsylvania was also recognized through the entire Salona and Coburn Formations and has been considered the GICE equivalent (Patzkowsky et al., 1997). In most cases the position of the GICE is constrained to the interval above the level of the Millbrig K-bentonite, which itself occurs within the Spechts Ferry Excursion immediately preceding the GICE. In platform and interior GACB settings, the GICE is typically developed as a two pronged excursion. The first part of the excursion represents the strongest and most pronounced part of the $\delta^{13}C$ curve and is coincident with the earliest part of the first major highstand of sea-level in the Chatfieldian and is demarcated by a higher total organic carbon component of the surrounding rocks (Fanton & Holmden, 2007). The second or upper spike of the excursion typically occurs a few meters higher in the late highstand to regressive component of the same sea-level cycle.

In West Virginia and Virginia, Young and colleagues (2004) recognized the major carbon isotopic excursion well above the position of the Millbrig K-bentonite (which occurs in the Nealmont Formation and Eggleston Formations) and above the V-7 K-bentonite. In fact, $\delta^{13}C$ values decreased following the Millbrig before they begin to climb into overlying rocks – a pattern that is typical in more interior regions as mentioned above. However this excursion is atypical in that it appears to be a much more protracted event that extends upward through the Dolly Ridge and "Trenton" Limestones before $\delta^{13}C$ values return to more typical values near the close of the Chatfieldian. In central Pennsylvania, Patzkowsky and colleagues (1997) recognized a very similar protracted event in the Salona and Coburn Formations that they had previously suggested was the "mid-Caradoc" or GICE equivalent excursion. However the anatomy and timing of this excursion relative to other nearby localities (including West Virginia, Virginia, and New York) is out of phase and is thus problematic. In all of the latter cases, the initiation of the GICE occurs nearly synchronously **above** the level of the Millbrig K-bentonite. In central Pennsylvania, in contrast, the calibration of the Patzkowsky et al. curve to the inferred position of the Millbrig K-bentonite in the Salona Formation shows the GICE to initiate well-**below** the level of the Millbrig. Thus either, 1) the GICE initiated in central Pennsylvania substantially earlier than any other location in the GACB or nearby regions thus far identified, or 2) the calibration of the GICE in Pennsylvania relative to the Millbrig K-bentonite is erroneous. Clearly, this needs to be considered in more detail; however, as suggested herein, it is more likely that the position of the Millbrig in central Pennsylvania is not yet established accurately, and its position is still to be discovered below the level of the GICE. Moreover, given the position of the Millbrig in the Nealmont Formation in West Virginia, it is likely that the Millbrig occurs within the Nealmont of Pennsylvania, as shown on **figure 9.**

Early Chatfieldian Neodymium Excursion

A second important chemostratigraphic system that has major implications for large-scale sea-level change and initiation of the largest tectophase of the Taconic Orogeny is recorded by neodymium isotopes. Discussed in more detail elsewhere herein, a number of papers (Fanton and Holmden, 2001, Fanton et al., 2002, Fanton & Holmden, 2007) have shown a number of excursions in neodymium isotopic composition of Chatfieldian-aged rocks from the Upper Mississippi Valley and in west-central Canada (**figure 10**). These publications have shown a major excursion in ε_{Nd} in sediments from the former regions suggesting a major change in sea-level and the supply of sediments to the GACB and transcontinental arch region. In these studies, changes in ε_{Nd} of sediments are thought to reflect changes in the delivery of sediments derived from old silicate rocks (more negative values) relative to the delivery of sediments derived from younger rocks (more positive values). In Laurentia during the mid to Late Ordovician, sediments were derived either from relatively old, craton-center rocks (i.e. Superior Province) or from relatively young, craton-margin rocks (i.e. Grenville Province). In most cases, sediments deposited in the vicinity of the Transcontinental Arch were derived from "old rocks" and thus have lower ε_{Nd} values. However in the Chatfieldian, during periods of sea-level highstand, these values periodically became more positive reflecting delivery of "younger"

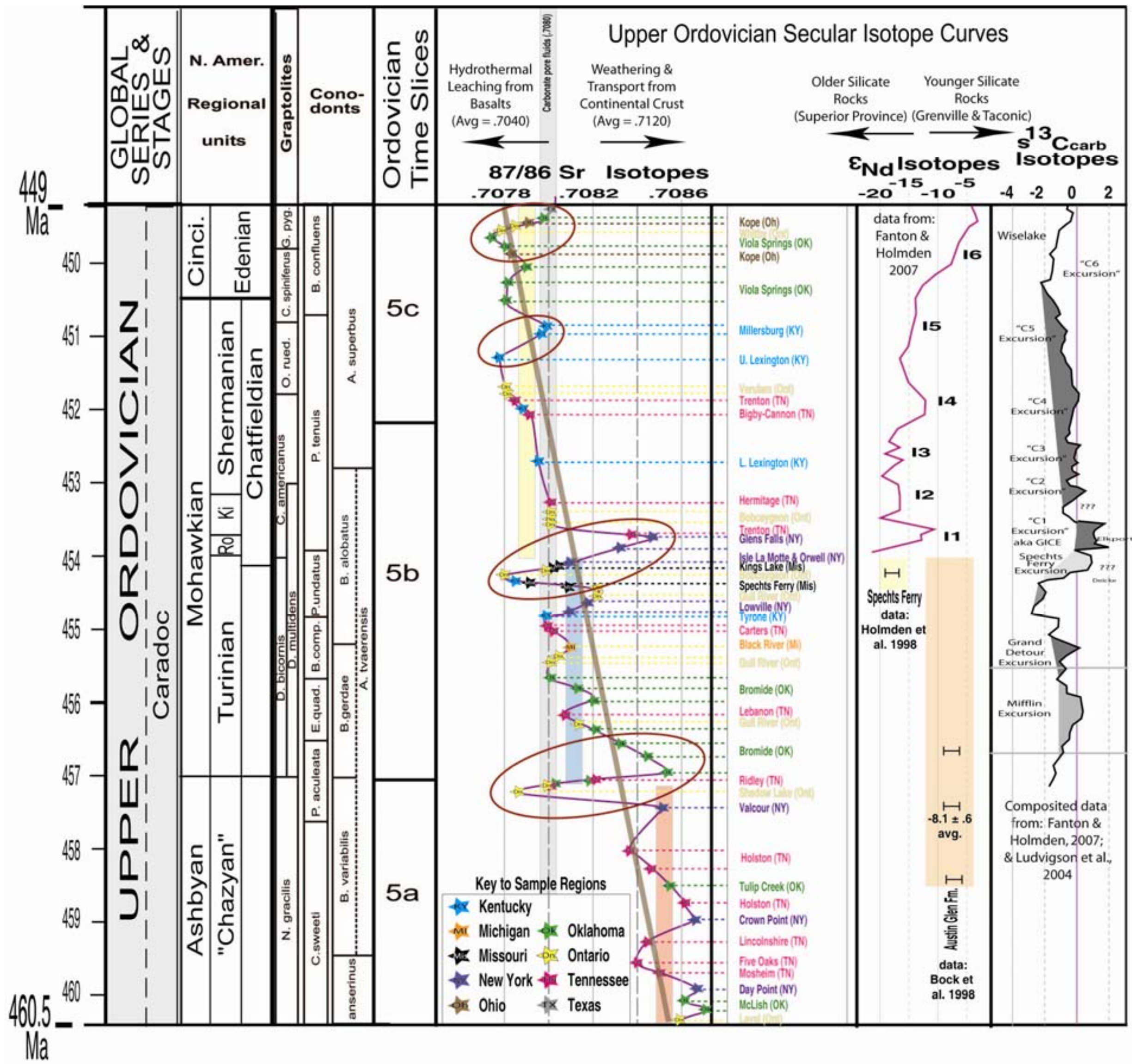


Figure 10: Secular Isotopic curves for the Upper Ordovician (Ashbyan to earliest Cincinnatian) of the GACB region. Included are strontium, neodymium, and carbon isotopic systems as integrated from previous studies.

sediments into the region – thus swamping the locally-derived sedimentary signature. In this unique case, ε_{Nd} acts essentially as a paleosea-level curve. Moreover, with much of the Laurentian craton submerged or nearly sub-merged during this time, the source of young silicate-derived sediments is inferred to have been from newly uplifted cratonic margin rocks, uplifted during the Taconic Orogeny.

As shown in **figure 10**, sediments deposited off the eastern coast of Laurentia during the Ashbyan to Turinian (Austin Glen Formation) show the elevated neodymium typical of Grenville-derived sediments (Bock et al., 1998), and generally maintain this relatively young provenance. In contrast, in the Upper Mississippi Valley, ε_{Nd} values are much lower leading into the Chatfieldian and then throughout the Chatfieldian to early Edenian gradually transition to Grenville signatures (Fanton and Holmden, 2007). However this long, drawn-out transition is punctuated overall by a number of short-term excursions that are shown in the figure. Especially important and prominent is excursion I1, recognized above the level of the Millbrig K-bentonite and coincident with the GICE.

The I1 excursion shows that sedimentation in the Upper Mississippi Valley was influenced for the first time by sediments derived from significantly younger sources. Fanton et al., suggest these are Grenville or cratonic-margin sediments that became remobilized during uplift and formation of the Taconic hinterland and were transported across the GACB region during the first major sea-level highstand. During this time much of the GACB was submerged and maintained a more-or-less uniform topography immediately prior to major faulting and development of craton interior shelf systems (i.e. the Lexington Platform, Galena Shelf, etc.) and the Sebree Trough. Subsequent excursions are recognized and correspond to depositional sequences (see chapter 7), but these are generally not as pronounced as the first I1 excursion perhaps due to development of major topographic barriers that would have blocked long-distance transportation of craton margin-derived sediments. Thus, the coincidence of the I1 ε_{Nd} excursion with the GICE appears to be a relatively robust and isochronous series of events.

Late Turinian to Early Chatfieldian Strontium Isotopic Excursion

During the transition from the upper Black River Group to the lowest Trenton Group, another important isotopic event is tentatively recognized for the first time herein. Using the stratigraphic framework of K-bentonite correlations, biostratigraphy, and patterns of lithostratigraphic change, previously published strontium isotopic compositions have been integrated to form a more highly-resolved, composited isotopic curve for the GACB region (**see figure 10**). As a result, several dramatic excursions are noted and discussed elsewhere herein (see chapter 6). However, a prominent paired excursion in $^{87}Sr/^{86}Sr$ values is recognized during the latest Black River to earliest Trenton. For reasons outlined elsewhere, changes in $^{87}Sr/^{86}Sr$ isotopic values are reflective of a number of processes that can be linked to major changes in sea-level and climate, and changes in rates of tectonic activity including volcanism and hydrothermal processes.

As shown in figure 10, $^{87}Sr/^{86}Sr$ values decline rapidly (more negative $^{87}Sr/^{86}Sr$ values) in the latest Turinian strata in the interval containing the Deicke and Millbrig K-bentonites. This is followed by a rapid increase in $^{87}Sr/^{86}Sr$ values increasing (more positive $^{87}Sr/^{86}Sr$ values) in the ensuing Chatfieldian. Constructed from data obtained from rocks in Ontario, Missouri, Kentucky, New York, and Tennessee, the second part of the excursion (the positive phase) is coincident with both the GICE and the I1 excursion. Thus this portion of the excursion occurs above the Millbrig K-bentonite, and appears to be present in multiple outcrop regions – although significantly more work has yet to be done to substantiate this initial analysis In this case, the coincidence of this event with a major carbon isotopic perturbation and a significant short-term shift in neodymium isotopes indicates that the Black River-Trenton interval likely records a significant perturbation in not only sea-level, but likely in terms of climate and tectonics as discussed elsewhere.

Although neodymium isotopic excursions in the upper Mississippi Valley are in themselves important in recognizing the delivery of siliciclastic sediments from the Taconic Orogeny, their timing is not recorded in the foreland basin setting (the region appears to only record the younger sediment signature). Therefore another isotopic system is needed to help corroborate the timing of the GICE event within the foreland. Although this analysis is only preliminary, it suggests changes in $^{87}Sr/^{86}Sr$ values at this time are coincident in a number of outcrop regions and thus might be another chemostratigraphic system that could be used to identify the timing of the GICE relative to K-bentonite horizons in central Pennsylvania.

SYNOPSIS OF BIOSTRATIGRAPHY & KEY EVENTS OF THE CBRT INTERVAL

As outlined above, there are a number of key observations that have been made here that build on previous studies and establish a number of possible and reasonable stratigraphic correlations across portions of the eastern GACB for this portion of the Late Ordovician. Salient observations documented include the following:

a) Herein, a compilation of previous biostratigraphic assessments for the entire Chazy, Black River, and Trenton groups and their equivalents from the principle study areas is made and are integrated with K-bentonite occurrences. Important contributions include:

- Strengthening the position of the *Phragmodus undatus – Plectodina tenuis* conodont biozone boundary and its position relative to the Turinian-Rocklandian Stage boundary. Recognized by Leslie (2000) in Tennessee at the base of the Hermitage Formation (2.9 meters above the Millbrig K-bentonite), and within the uppermost Chaumont Formation of

Ontario/Quebec (Barnes, 1967) also above the Millbrig K-bentonite, the biozone boundary is now confidently located below the base of the Trenton Group and below the historical position of the Rocklandian Stage boundary in the type region. The absence of *P. tenuis* in deep-water facies in Kentucky, Pennsylvania, West Virginia, and elsewhere prevents recognition of the basal biozone boundary in these regions, although it is approximately coeval with the position of the Millbrig K-bentonite.

- The Deicke and Millbrig K-bentonites are located within, but near the top of the *Diplograptus bicornis* graptolite biozone (Webby et al., 2004). In Pennsylvania, graptolites characteristic of the *Corynoides americanus* graptolite zone (immediately younger than the *D. bicornis* zone) are associated with the lowermost K-bentonites of the Salona Formation. As two of these K-bentonites have previously been correlated with the Deicke and Millbrig respectively, graptolite biostratigraphic zonation is in disagreement and suggests that the K-bentonite correlations are not accurate. Graptolite biozonation in central Pennsylvania suggests the basal Salona is younger than had previously been assumed using K-bentonites.
- With respect to recognition of the Deicke and Millbrig K-bentonites in New York. The Millbrig has been correlated with the Hounsfield K-bentonite of New York State (Mitchell et al., 2004). As discussed herein, the stratigraphic nomenclature employed by these workers, indicated that the Millbrig occurred within the base of the Selby Formation and therefore at the base of the Rocklandian Stage. Based on the result of detailed stratigraphic analysis and correlation of local rock units between more complete sections, it is suggested here (& see chapter 3) that the Hounsfield/Millbrig actually occurs in the

Glenburnie Member just below the base of the Watertown Member of the Chaumont Formation and some 4-5 meters below the level of the Selby Limestone.

- Furthermore, a K-bentonite does occur at the base of the Selby Formation; however, it is not exposed in the Hounsfield type-locality. This K-bentonite has tentatively been assigned a Deicke K-bentonite signature (see Carey, 2007). As the Deicke is always below the Millbrig, this higher K-bentonite cannot be the Deicke K-bentonite as documented elsewhere in the GACB.
- In terms of macrofaunal analysis, a framework of epiboles and recurrent faunal/floral associations has been established into a series of recurrent zones.
    i. Typically these recurrent zones are characterized by the return of a preferred environment and water depth for the faunal association during deepening and highstand episodes. In the Chazy-Black River, the recurrent high diversity zones are characterized by protected open-marine associations dominated by larger colonial corals, horn corals, stromatoporoids, and a diversity of bryozoans, brachiopods, bivalves, and other taxa.
    ii. In the Chazy-Black River, recurrent zones are distinctly separated from one another by intervals composed of more restricted facies characterized by low diversity faunas/floras. These often are composed of stromatolitic/algal mat associations with endemic bryozoan, gastropod, and ostracod faunas that are contained within mud-cracked, fenestral micrite facies often containing smaller colonial (*Tetradium*) corals.
- Major lithostratigraphic transitions within the CBRT interval show evidence for associated local origination and extinction events.

i. In pre-Trenton shallow-water strata, trilobite associations included bathyurids as dominant forms. The rapid loss of shallow-water environments at the end of the Turinian (to very earliest Rocklandian) and the coeval onset of major shale deposition across the GACB is coincident with the extinction of this group.

ii. The former extinction event is coincident with a major extinction in cephalopod faunas that show a pronounced drop in diversity at about the level of the Millbrig K-bentonite (Frey, 1995).

- The onset of the first major shale deposition, and associated widespread deepening, across the GACB in the Rocklandian Stage, is coincident with the incursion of deep-water, cryptolithid trilobite taxa and *Prasopora* bryozoans into eastern Laurentia.

i. Cryptolithids are first found in most outcrop areas within the deepest water facies of the Rocklandian stage. They are argued to have invaded the eastern GACB from the Northwest Territories, possibly through the Timiskaming Strait of eastern Ontario-Western Quebec, once sea-level reached a critical depth. Thereafter, their recurrence/migration in the eastern GACB and component sub-basins/platform areas is controlled by major deepening events. As a group, they changed very little throughout the Chatfieldian until the Edenian when endemism (?) resulted in a much more diverse range of forms.

ii. In the Trenton *Prasopora* are facies specific and have characteristically high abundances when the preferred facies (relatively deep-water, mixed siliciclastic-carbonate association) is best developed. *Prasopora* likely immigrated into the eastern GACB region, by way of the Upper Mississippi Valley, from Oklahoma as siliciclastic-influenced deposition initiated in the latest Turinian near the

Transcontinental Arch (uppermost Platteville Formation to lower Decorah Formation). Its first major widespread epibole across the platform was approximately coincident with the incursion of cryptolithid trilobites in the early Chatfieldian (Rocklandian). Thereafter, *Prasopora* is abundant in muddy facies of the Taconic Foreland and Sebree Trough where it underwent species splitting events in mid to late Chatfieldian time. *Prasopora* become abundant in cratonic interior sub-basins and on platform areas during highstand events and therefore are often useful for identifying and correlating the deepest facies of depositional sequences (see chapter 7).

- A second trilobite invasion is noted through the first incursion of *Triarthrus* in middle Shermanian (mid Chatfieldian) time in the Taconic Foreland Basin during another substantial deepening event associated with an expansion in low-oxygen conditions.
    i. Again an immigrant from northwest of the Transcontinental Arch (Alaska) this form appears in the Taconic Foreland basin in deepwater facies in New York, and Pennsylvania at this time.
    ii. The form also makes synchronous incursions into craton interior areas including the Sebree Trough during the earliest Edenian Stage. As with *Cryptolithus, Triarthrus* shows evidence for speciation in the Edenian leading to a number of different forms.
- Ostracod biostratigraphy is most valuable in the Ashbyan to very earliest Turinian. Thereafter ostracods show a high degree of endemism in local outcrop areas, or are long ranging taxa that are not useful for biostratigraphy.

i. Two important biozones (*Bullatella kaufmanensis* & *Monoceratella teres*) have been recognized by Swain (1957).

ii. These help in correlation of the Chazy Group and its lateral equivalents in Pennsylvania, Virginia, and Oklahoma.

- Six major echinoderm assemblages have been recognized in the Chazy, Black River, and early Trenton interval of eastern Laurentia.

  i. These include the Crown Point, Benbolt, Lebanon, Platteville, Curdsville, and Kirkfield assemblages. Additional associations are being recognized now in the middle to upper Trenton as well. These associations have been variously correlated and some have been considered synonymous and synchronous (i.e. Curdsville and Kirkfield).

  ii. Herein it is suggested that some of these associations represent widely correlative, yet distinct intervals that were developed when environmental conditions were optimized for echinoderm growth and diversification. In the Ashbyan-Turinian, these generally occurred during periods of sea-level highstand. Subsequently during the Chatfieldian, when siliciclastic influx became significant across much of the GACB, echinoderm–rich associations become more abundant during sea-level rise when siliciclastics are starved.

  iii. During the Ashbyan to early Turinian interval, echinoderm-rich associations were dominantly restricted to platform margin areas. However after an early Turinian restriction event, beginning with the Lebanon assemblage, diverse echinoderm associations are recognized in coeval facies across much of the GACB, including in platform interior areas (i.e. upper Platteville Formation of the Upper Mississippi Valley).

- An increasing number of brachiospongids (hexactinellid sponges) are now being recognized in Chatfieldian strata, and it now appears that these forms evidently evolved and diversified in the early Chatfieldian of eastern Laurentia.
    i. Two major epiboles are recognized: 1) early Chatfieldian; Curdsville (Kentucky), Hermitage (Tennessee) and lower Bobcaygeon Limestones (Ontario) (*Brachiospongia tuberculata* with 7-8 arms), and 2) mid to late Chatfieldian, Brannon (Kentucky) and upper Verulam to lower Lindsay Limestones (Ontario) (*B. digitata* with 9-10 arms).
    ii. Two additional forms are known from intervening strata in Ontario. These have not yet been recognized elsewhere.
- In addition to Prasopora mentioned above the bryozoan *Constellaria* is also important for correlation across the GACB region.
    i. *Constellaria* is well known from particular levels in the Cincinnatian, but recurrent zones occur at different levels and are useful for correlations in underlying strata as well.
    ii. These include: 1) the late Ashbyan of New York (Chazy) and equivalents to the south (including the Pierce and Ridley Formations of Tennessee), and the 2) mid-Chatfieldian of Kentucky, Tennessee, New York, and possibly Pennsylvania.

b) In terms of K-bentonite-based event stratigraphy, this study highlights the controversies that remain with regard to correlation of major K-bentonites (especially the Deicke and Millbrig) as were previously documented by Haynes (1992) and Kolata and colleagues

(1996). As discussed above, the Deicke and Millbrig K-bentonites have been the focus of a number of chemical fingerprinting studies and, especially in New York and Pennsylvania, the results have been problematic and are in disagreement with other correlation assessments.

- In New York, it is suggested, that the Hounsfield/Millbrig K-bentonite position reflect the stratigraphic assessment first proposed by Kay (1931) rather than the revised assessment of 1935.
    i. Given the correlations established here (and see chapter 3) the lithostratigraphic context of the local region suggests the 1931 assessment is more accurate.
    ii. Using the former assessment, chemical fingerprinting of the Hounsfield by Mitchell and colleagues (2004) establishes the position of the Millbrig K-bentonite below the top of the Watertown Member of the Chaumont Formation (and not above it).
    iii. Therefore the Millbrig occurs below the top of the Turinian Stage and a short distance below the base of the Rocklandian Stage (~5 meters).
- With regard to central Pennsylvania, it is important for K-bentonites below the level of the Salona Formation (N1, N2 and N3 K-bentonites) be collected and fingerprinted for comparison to Deicke and Millbrig signatures.
    i. It is increasingly unlikely that the Salona K-bentonites are equivalents of the Deicke and Millbrig elsewhere.
    ii. This assessment is based on the initiation of the Guttenberg Isotopic Carbon Excursion (GICE), lithostratigraphic assessments, and various biostratigraphic data as previously discussed.

- Possible resolutions and suggestions for future analyses have been identified herein. In addition, the possibility has been raised that multiple K-bentonite horizons might contain similar chemistries, even though they are not time-equivalent.
- Alternatively, given the number of erosional surfaces recognized in the late Turinian to early Chatfieldian, it is possible that K-bentonites can be remobilized and re-deposited after their first episode of deposition.

c) Previous chemostratigraphic event studies from multiple areas across the GACB region have been integrated herein (and see especially chapter 6).

- As indicated by Ludvigson and Witzke (2005) and Fanton and Holmden (2007) there are an increasing number of carbon isotopic excursions in the Platteville – Dunleith formations. These additional excursions complicate their use in correlation studies. Nonetheless a significant positive isotopic carbon excursion has been recognized across the GACB by former workers.
    i. The Guttenberg Isotopic Carbon Excursion (GICE) is consistently recognized and identified in Iowa, Minnesota, Kentucky, Virginia, West Virginia, and New York at a level some distance above the Millbrig K-bentonite (see Young et al. 2005) and is generally most-well developed between the Elkport and Dickeyville K-bentonites and their equivalents.
    ii. As documented by Fanton and Holmden (2007) the GICE has been shown to be synchronous with a neodymium isotopic event (I-1) that has been interpreted as reflecting a significant deepening phase and the first major pulse of siliciclastics from the Taconic Orogeny. Herein it is shown that this event occurs within the Rocklandian and is coincident with many of the faunal epiboles discussed above.

iii. In the Taconic Foreland Basin in central Pennsylvania and West Virginia, the GICE is anomalous in that its duration is evidently expanded upward into the mid to late Chatfieldian (see Patzkowsky et al., 1997; Young et al., 2005) whereas on the interior platform it is restricted to the early Chatfieldian (Rocklandian to lowermost Kirkfieldian). Herein it is suggested that the "expanded" excursion may actually represent a composited excursion composed of multiple carbon isotopic excursions as now recognized in the Upper Mississippi Valley by Fanton and Holmden (2007).

- Neodymium isotopic studies show that siliciclastics in the Chazy-Black River interval of the eastern United States show relatively young sources for neodymium isotopes (Grenville & Taconic; as per Bock et al., 1998).
  i. In contrast, Fanton and Holmden (2007) show that neodymium in the upper Mississippi Valley are much older during this same time (sediments likely derived locally from the Superior Province/Transcontinental Arch).
  ii. Beginning with the GICE event, and during subsequent flooding events, neodymium isotopes in the Upper Mississippi Valley show the same young "Grenville-Taconic" source as identified in the east by Bock et al.
  iii. This suggests siliciclastics were transported across the entire GACB from source areas in the east, i.e. the newly uplifted thrust sheets of the Taconic accretionary prism during the first significant deepening coincident with deposition of the first Trenton Group sediments.
- Previous strontium isotopic studies (see chapter 6) have shown an extended period of decline in $^{87}Sr/^{86}Sr$ ratios through the late Ordovician. Based on biostratigraphic,

lithostratigraphic, and other event correlations discussed herein, it has been possible to integrate data from multiple previous studies into a single, chronological, secular curve.

i. This curve shows that the long-term decline in $^{87}Sr/^{86}Sr$ ratios is punctuated by several paired excursions – each with a sharp negative followed by sharp positive excursion.

ii. The second of these "paired excursions" is situated with the major lithostratigraphic and biostratigraphic transition that demarcates the Turinian-Rocklandian boundary interval (Black River-Trenton Group contact interval).

iii. A rapid decline in the $^{87}Sr/^{86}Sr$ ratios is noted in the interval containing the Deicke and Millbrig K-bentonites (latest Turinian). Given the volume and intensity of volcanism at this time, this decline is thought to be associated with an increased volume of strontium leached from hydrothermal alteration of basalts on the Late Ordovician seafloor.

iv. This decline is followed by a rapid increase in $^{87}Sr/^{86}Sr$ ratios going into the early Chatfieldian coincident with both the GICE and the I1 neodymium excursion. This positive shift is likely the result of increased weathering and transport of siliciclastics from continental crust (and/or Taconic-derived sediments) during the initiation of the main phase of the Taconic Orogeny.

- Collectively, these isotopic excursions, all within the latest Turinian to early Chatfieldian (Rocklandian) indicate that the Late Ordovician oceans, the GACB epicontinental sea, and the fauna/flora therein were substantially influenced by a rather rapid, synchronous shift in sediment supply, sea-level, tectonic regime, and likely climate change. During the ensuing Chatfieldian, the GACB witnessed major changes in basin-platform architecture.

CHAPTER 3: Summary of Detailed Stratigraphy of the Ashbyan to Mohawkian Interval in its Type Area: New York and Ontario

ABSTRACT

This chapter provides a detailed sedimentary and stratigraphic description of the Chazy, Black River and Trenton Groups, and the paleogeographic context of their deposition in the New York to eastern Ontario region. The discussion is intended to provide an integrated synopsis of previous work and as an update of New York stratigraphic nomenclature for the entire Ashbyan to Mohawkian interval of the Late Ordovician. Moreover, this stratigraphic discussion provides a stratigraphic base-line model with which the stratigraphic successions of other regions of the Great American Carbonate Bank (GACB) are compared. Important new contributions in this study include detailed descriptions of the Chazy-Black River group boundary interval, as well as that of the upper Black River – to lower Trenton Group interval.

INTRODUCTION AND HISTORY OF STUDY:

The Black River and Trenton Groups have been studied for well over a century and have been, almost since the beginning, a topic of debate resulting in many revisions and reassessments in stratigraphic nomenclature. Part of this contention stems from: 1) the predominance of these units in early geologic studies of North America such that these units were often used as text-book examples for the education of students, 2) the implementation of new stratigraphic methods (using advanced modern stratigraphic techniques) which required the re-evaluation of some of the older long-standing ideas, and to some extent 3) the complex and disconnected outcrop regions which offer minimal continuous exposure and apparent geographic variability.

Despite the long-standing history of research, the Black River and Trenton Groups are once again the subject of major new discoveries, especially in oil and gas within the subsurface of New York, Pennsylvania, West Virginia, and within the larger cratonic setting of eastern Laurentia. Moreover, newly recognized chemostratigraphic events have been studied over fairly broad regions of eastern North America and even in Scandinavia and Estonia (Ludvigsen et al., 1996; 2004; Saltzman et al., 2001; Bergström et al., 2004) and suggest that the transition from the Black River into Trenton was a fairly major event that certainly had implications in the trans-Iapetan region and may have even been global.

STRATIGRAPHY OF THE CHAZY GROUP

<u>General Description, Distribution, and Members of the Chazy Group</u>

The earliest Late Ordovician Chazy Group was deposited during the onset of the Tippecanoe Megasequence (Landing & Westrop, 2006). The type section **(figure 1)** is located in

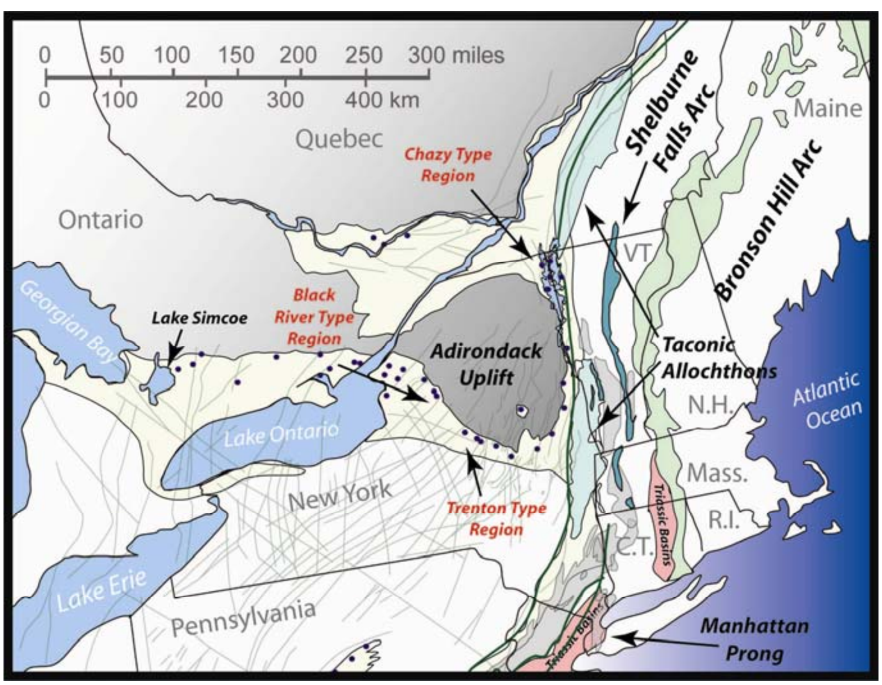


Figure 1: Ordovician outcrop areas (yellow shading) of the Northeastern United States showing the type areas for the Chazy, Black River, and Trenton Groups as well as key structural features of the region including the Adirondack Uplift and the Frontenac Arch that connects the Adirondacks with the Canadian Shield. The Adirondack Uplift separates the

Black River and Trenton type-regions from the Chazy-type region and was likely at least intermittently influential in deposition during the Ordovician.

the Champlain Valley of New York and Vermont as well as in the adjacent St. Lawrence River Valley of easternmost Ontario and southern Quebec, Canada. The Chazy is limited in its distribution in this region to areas east of the Frontenac Arch and east of the Ottawa Embayment (**Figure 2**). For the New York region the limited distribution of Ashbyan age rocks is likely

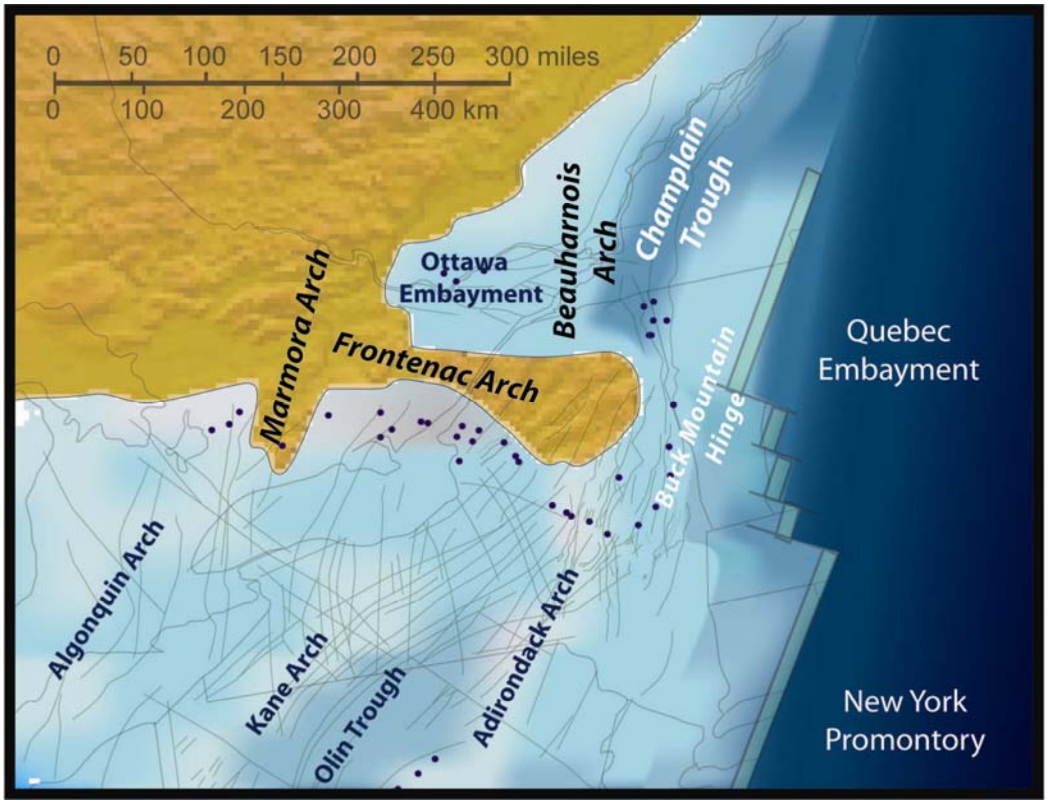


Figure 2: Paleogeographic interpretation for the type region of the Chazy, Black River, and Trenton Groups during the Late Ordovician prior to the major onset of Taconic orogenesis –during the greatest extent of the Black River. Key regional features are located as recognized from sedimentologic and stratigraphic evidence.

due to two reasons. First, during the onset of the Tippecanoe transgression, sea-level had only barely risen onto the eastern margin of Laurentia. The result is that the Chazy Sea in this region had a limited lateral extent. Second, based on studies (Fisher, 1968) of the Beekmantown Group (underlying the Chazy), there is significant evidence for development of the Adirondack Arch during the pre-Tippecanoe sea-level lowstand. The arch is identified by the uplift and truncation of rocks underlying the Chazy and maybe significant of the initiation of tectonic collision off the coast of Laurentia (Jacobi; 1981). As such, the region may have become slightly uplifted (along a peripheral bulge) so that Chazy deposits, if deposited elsewhere, may have been eroded subsequent to deposition of the overlying Black River Group.

Despite the limited exposure and areal distribution, Chazy rocks of the eastern New York State region show a number of lithologic and paleontologic characteristics that help to ascertain the depositional environments in the Chazy Sea. First the region supported abundant, diverse open marine fauna characteristic of Boucot's (1975) benthic assemblages (BA-1 through BA-3). Characteristic of these assemblages are some of the first stromatoporoids, several new species of brachiopods, some of the first skeletonized corals (and some of the first reefs). Also represented are specialized nautiloids, trilobites, many mollusks including the key index fossil: *Maclurites magnus* Le Seur. Also represented is a diverse assemblage of echinoderms and many new bryozoans, who also became important reef builders in this region (Isachsen, 1991).

The Chazy Group, as reported by Oxley and Kay (1959), is estimated to be about 225 meters (740 feet) thick in the type section (although even there it is an incomplete exposure) (**figure 3**). However its thickness does vary in the region. The Chazy Group increases in thickness toward the north into Quebec and to the eastern side of Lake Champlain before thinning again across the Buck Mountain Hinge of Welby (1961). The Chazy also thins westward and southward where the Chazy onlaps Precambrian Grenville basement rocks near the Adirondack Dome. Here thicknesses are less than 90 meters, but eventually the Chazy pinches out at the base of the Black River Group. Both Welby (1961) and Fisher (1968) implied that some of the thickening and thinning may be related to faulting and/or local paleotopographic variations at the time of deposition. Based on thickness and lithologic variations in the Chazy, it is likely that the modern Champlain Basin is roughly centered in the position of a localized (fault

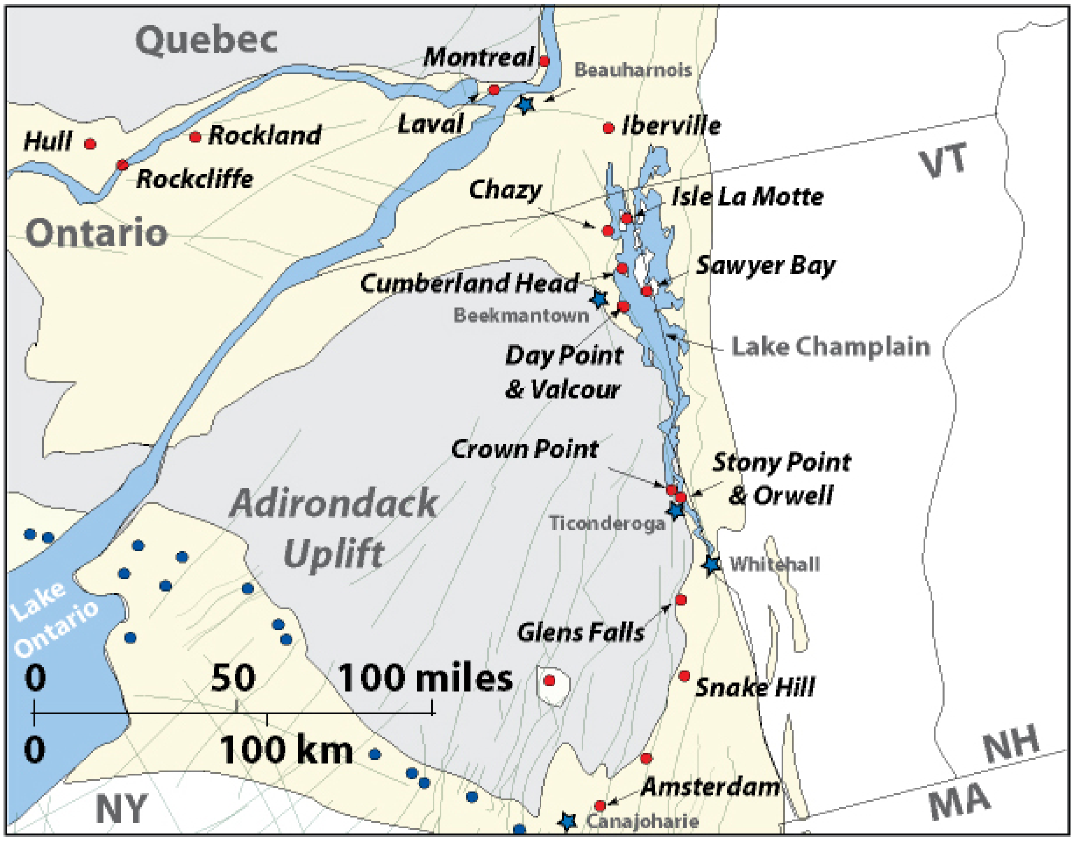


Figure 3: Type-section localities for Chazy, Black River, and Trenton Group localities for the Lake Champlain to St. Lawrence Seaway Region of northern New York, Southeastern Ontario and southern Quebec. Shown in yellow is the distribution of Ordovician-aged rocks, and in grey are Precambrian-Cambrian aged rocks.

controlled?) sub-basin of the Ordovician GACB. The axis of the Ashbyan aged sub-basin apparently ran approximately north-south. It was bordered by, and parallel to, the Beauharnois Arch on the west and the "Buck Mountain Hinge," of Welby (1961) on the east (**see figure 2**).

Although Chazy deposits crop out near Ticonderoga and Crown Point, the southernmost limit of the Chazy sub-basin is not well-exposed, nor well-defined, as these rocks disappear in the vicinity of the Taconic over-thrusts. Fisher (1968) shows these units pinching out in the vicinity of Whitehall, New York by both basal onlap of the Adirondack Arch to the south and by erosional top truncation in that direction at the base of the Black River Group. He does not show them occurring in the Mohawk Valley region (another 20 miles south of Glens Falls). Selleck and Baldwin (1985) suggested that Chazy equivalents were however present east of Whitehall (east of the Buck Mountain High) beneath the Taconic thrusts as some small slivers of these fossiliferous rocks were included in the basal thrusts. Equivalent rocks are exposed in the Middlebury Syncline and are referred to as the Middlebury Limestones although they have been deformed and approach marble.

With increased thicknesses generally in the northern Champlain area, it is apparent that that region experienced more subsidence and presumably made connection with deeper water through the Quebec reentrant to the northeast. If Fisher's southern limit of the Chazy is correct, there appears to have been an uplifted area in the vicinity of Glens Falls that precluded deposition in that region. This high may be the north-eastern extent of the Canajoharie Arch of Fisher (1980) or the Adirondack Arch of Kay (1935). This feature appears to have been a positive topographic feature extending above sea-level and was exposed repeatedly throughout the Ashbyan to Turinian before becoming submerged by sea-level rise, and or tectonically down-dropped during the deposition of the Trenton Group.

The Chazy Group is comprised of three formations, and in ascending order these are the: Day Point, Crown Point, and Valcour Formations (Cushing, 1905). The basal contact of the Chazy with the underlying Lower Ordovician Beekmantown Group has been described as a fairly subtle one in the type region. Perceived as a slight hiatus or break by Welby (1961), he suggested that there was little evidence for an angular unconformity in the Champlain Valley region. Nonetheless work by Rickard (1973) and Fisher (1980) suggested a significant amount of truncation for the top of the Beekmantown and equivalents in the eastern Mohawk Valley. They surmised a similar truncation in the Chazy area. Nonetheless, Welby considered the contact as paraconformable and recognized the contact on a lithologic basis by the change from carbonates (of the underlying Beekmantown) to quartz-dominated rocks of the basal Day Point.

Without well-preserved fossils in the basal Day Point and underlying Providence Island Dolostone (uppermost Beekmantown) it was difficult to ascertain the amount of hiatus. Subsequent paleontologic work, using trilobite and conodont biostratigraphy (Landing et al., 2003; Landing & Westrop, in press), help to establish a fairly significant time-gap at the top of the Beekmantown in the Mohawk Valley (spans the entire upper Lower Ordovician and Middle Ordovician). In the Champlain Valley, the gap is significantly less with the Middle Ordovician Providence Island Formation of the uppermost Beekmantown underlying the Chazy Group (Landing & Westrop, in press). Thus it appears that Welby's (1961) assumptions were more accurate than those of Fisher (1968) at least in this region. Nonetheless, this hiatus, whether in the eastern Mohawk Valley or the Lake Champlain region, is synonymous with that recognized by Sloss (1963) as the Knox Unconformity and the base of his Tippecanoe Megasequence.

The upper contact of the Chazy Group with the overlying Black River Group is equally enigmatic. Described as a regional disconformity (Selleck and Baldwin, 1985), the upper member of the Chazy (the Valcour) is also documented to grade cyclically upward into the overlying Black River. This transition is recorded by a series of buff-weathering argillaceous dolostones that intercalate with light gray colored calcilutites typical of the Black River. Fisher (1968) suggested that the break was not obvious in the Chazy region despite "unrivaled" exposures. In his analysis he depicted no gap in sedimentation in the northern Champlain Valley region between the Chazy and overlying Black River. This was also suggested by Johnsen and Toung (1960) who indicated that these rocks are not distinguishable lithologically from the basal Black River member (the type Pamelia). Farther south in the Champlain Valley, the Valcour becomes much thinner and is eventually truncated from the top as shown by Rickard (1973).

Chazy Group Formations: Day Point

The lowest Upper Ordovician Day Point Formation ranges in thickness from ~ 25-90 meters (80-300 feet) in thickness. The type section is located southeast of Beekmantown, New York on the southern end of Valcour Island in Lake Champlain– just opposite Day Point on the western shore. Cushing (1905) formalized the Day Point Formation after Brainerd and Seely (1896). Subsequently, Oxley and Kay (1959) formally described four internal members (Head, Scott, Wait, and Fleury) as recognized by Brainerd and Seely (1896) on the islands of Valcour and Isle La Motte. Outside of these localities in northern Lake Champlain, subdivision of the Day Point into members has met with some difficulty due to rapid lateral facies change across local topographic features and onlap patterns toward the central and southern Champlain Valley exposures (Selleck and Brewster, 1985).

In the type area the basal unit, the Head Member, rests on the Providence Island Dolostone. It is comprised of dark-gray, greenish-tinted, cross-bedded quartz sandstones and coarse siltstones with minor shale interbeds (Oxley & Kay, 1959). The Head Member is typically about five meters thick and displays occasional *Lingula* sp. In Vermont, the unit becomes dominated by coarse siltstones and is referred to as the St. Therese Siltstone (Fisher, 1977). Up-section lithologies grade into quartz-rich, skeletal-carbonates of the Scott Member that carries a restricted fauna including bryozoans, ostracodes, trilobites, and *Lingula brainerdi*. Overlying the Scott is a relatively thin interval of fossiliferous quartz sandstone to coarse siltstone, again with *Lingula brainerdi*. This unit is the Wait Member and apparently has a localized distribution (northwestern-most localities). The topmost member of the Day Point Formation is the Fleury Member. It ranges between 18 to 61 meters thick (60-200 feet). It is characterized by massive beds of planar to cross-bedded echinoderm-rich calcarenites with minor calcareous shale interbeds (Shaw, 1969). Although usually medium-light grey, in weathered sections, this unit often takes on reddish coloration especially close to the top of the section. In the Plattsburgh area, this reddish, iron-stained interval has been quarried for building stone and has been referred to as "Lepanto Marble" (Fisher, 1968).

In addition to its transition upward out of siliciclastics into coarser grained carbonates, the Day Point contains an increasing abundance and diversity of taxa. It displays a number of bryozoan and echinoderm bioherms or reefs. In fact this member shows domal bryozoan-coral-sponge-algal reefs that are generally smaller than six meters in diameter. These are often found within fifteen meters of the contact of the overlying Crown Point Formation. In the calcarenites that flank these reefs, brachiopods are extremely abundant (Fisher, 1968). These sediments likely were deposited in shallow, relatively high-energy environments. Collectively, lithologic and paleontologic data suggest an overall deepening-upward or transgressive pattern within the Day Point Formation (Shaw, 1969).

Chazy Group Formations: Crown Point

The middle unit, the Crown Point Formation, is the most widespread unit within the group and extends furthest to the south along the Champlain Valley before it goes to extinction (**figure 4**). It ranges in thickness from 15 to over 75 meters. The Crown Point is typified by argillaceous medium to fine-grained limestone ranging in color from black to a medium grey (Welby, 1961). In some instances this lithology appears to be sub-lithographic with a finer-grained composition. Individual layers range from about five to fifty centimeters in thickness, (thin to thick-bedded), with thicker beds showing significant amounts of bioturbation resulting in the amalgamation of several thinner beds. Among the dominant features of this formation are

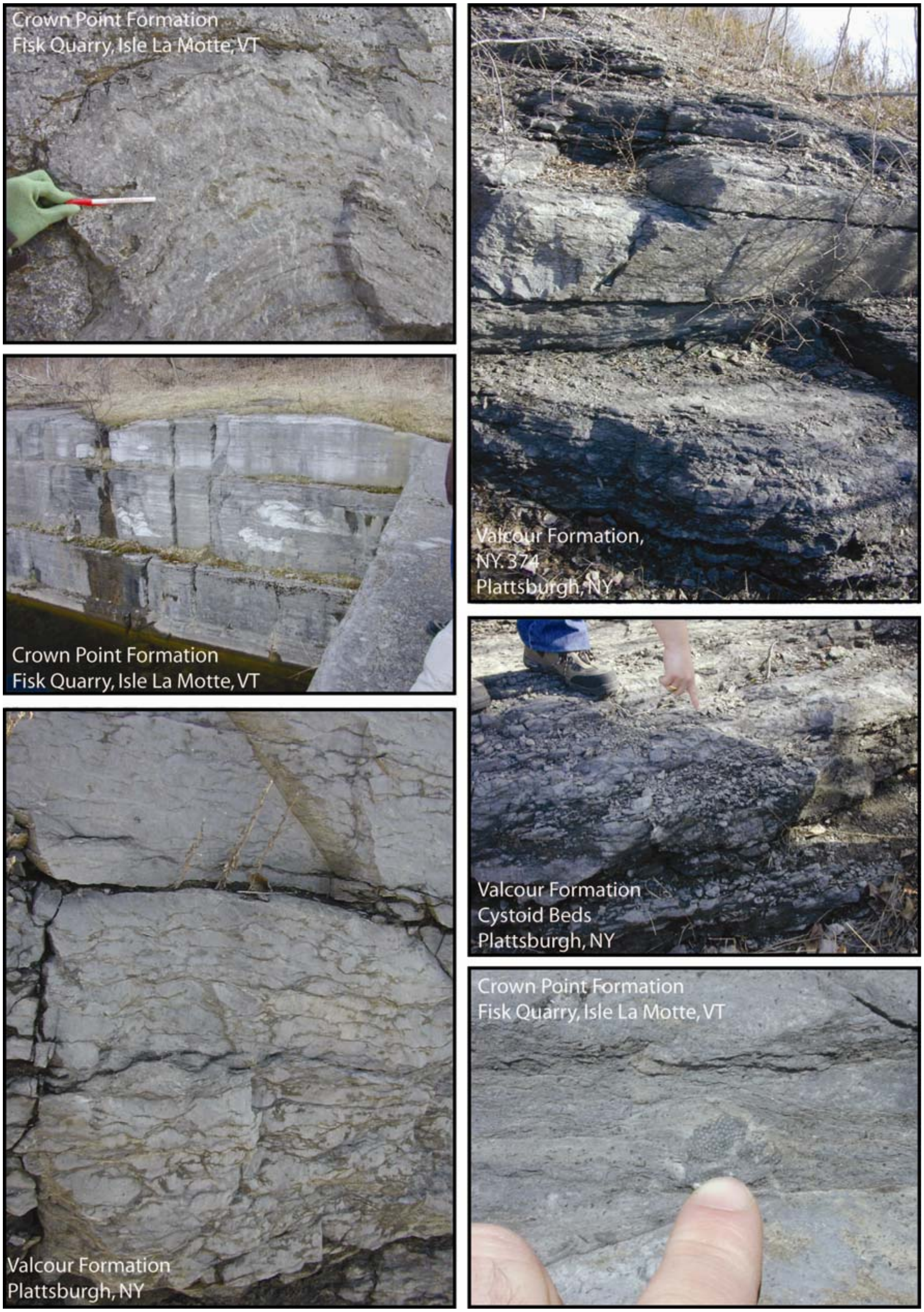


Figure 4: Outcrop photographs of the Crown Point and Valcour Formations from the northern Champlain Valley region. Outcrops in the Fisk Quarry, Isle LaMotte, VT, show the development of extensive Stromatoporoid biostromes and associated reef facies. Also shown are sections of the Valcour Formation north of Plattsburgh, New York, including the beds yielding specimens of an as yet undetermined cystoid.

the thin (1-2 cm), cross-bedded and medium to coarse-grained dolomitic stringers interspersed within the bioturbated fabric and infilling burrows (Shaw, 1969). In most sections, shale partings are rare (Fisher, 1968) except for some silty-shale beds near the base of the formation near the contact with the underlying Day Point.

Although dominantly carbonate the Crown Point, like much of the Chazy, also has a quartz sandstone component. Near the margins of the Chazy outcrop limit, and within the upper portion of the Crown Point, are occasional lenses composed of sub-angular to sub-rounded quartz grains. In many cases, quartz-grains often form the cores of small centimeter-sized *Girvanella* blue-green algal nodules that are common throughout the formation. In addition to quartz, another indicator of the Crown Point interval is the occurrence of blue-black chert nodules. Welby (1961) describes them as being more prevalent toward the summit of the formation although they are also found at the base in the type area.

The Crown Point is known for the appearance of small bioherms with some of the first stromatoporoids (*Cystostroma vermontense* and *Stramatocerium sp.*) and several larger colonial corals species as well as a solitary rugose species. Unlike the underlying Day Point reefs, these structures are composed of fine-grained carbonates and lack coarse-grained flanking calcarenites. Thus the major difference in lithology between the Crown Point and Day Point, and even the overlying Valcour Formation, is the lack of coarser-grained limestones (Fisher, 1968). Shaw (1969) considers it likely that the Crown Point Formation, on the basis of its lithologic and paleontologic descriptions, was deposited in deeper less agitated waters than the Day Point Formation (Shaw, 1969).

Chazy Group Formations: Valcour

The Crown Point Formation grades upward into the highest unit in the Chazy Group. The uppermost formation, the Valcour, ranges in thickness but is estimated to be between 33 to over 60 meters (110 -200 feet) in thickness in the northern Champlain Valley. It thins to the west and pinches out to the south within the central Champlain Valley area. It is comprised of two members, the lower Hero, and the upper Beech (Shaw, 1969). Although not easily identifiable in all exposures, the members usually help to differentiate the unit from the underlying Crown Point. This unit is the most argillaceous of the Chazy Formations and is dominated by interbedded dolomitic limestones and occasional dolostones. Many of the dolomitic intervals appear in intensely bioturbated coarser-grained fabrics. The lower 12 to 17 meters of the unit contain reefs constructed from bryozoans, sponges, algae, and stromatoporoids (as in the underlying Crown Point). However, these reefs are often associated with and interbedded with medium-light grey calcarenites that in some areas show evidence for channeling and much shallower water deposition signified by occasional herringbone cross-stratification. The “channels” in shallow reefs are commonly filled with trilobite and nautiloid fragments and carbonate sands. Collectively these data imply a return to shallow water (Shaw, 1969).

The remainder of the Valcour formation, the upper or Beech Member is demarcated in fresh surfaces by darker grey, argillaceous calcilutites with shale partings, argillaceous dolostones that commonly show evidence for restricted faunas, and greenish-gray quartz siltstones and sandstones (Cooper, 1956, Welby, 1961; Fisher, 1968). These beds also show evidence for desiccation features and cryptalgal lamination. On weathered outcrops, these rocks commonly show yellow-orange or buff weathering surfaces. The uppermost beds of the Valcour show an interesting transition from dolomitic limestones into peritidal birdseye micrites similar to those of the Lowville Formation (Black River Group). In Quebec the transition zone is referred to as the Pamelia Formation (Black River Group) and is known for some thin sandstone beds near the top.

In the Champlain Valley region the boundary is not well established and has been referred to as transitional (Fisher, 1968). Recent work in Quebec (Salad Hersi & Lavoie, 2001), document the boundary contact as an abrupt change from the Laval dolostones (Chazy-equivalent) into the basal Pamelia sandstones (or Lowville where Pamelia is absent). In the Montreal to Quebec region, this contact is seen as unconformable with overlying Black River Group units showing a renewed onlap over and beyond the extent of formerly deposited Chazy. In the type Chazy area the sandstone units as recognized by Salad Hersi and Lavoie (2001) are not as well developed and the contact is significantly less apparent. Although more work needs to be done to confirm it, it is suggested that the Valcour as previously defined likely contains the same erosional unconformity. Thus the uppermost Valcour represents an equivalent of the Lower Pamelia of the type region in western New York State as well as the Pamelia of the St. Lawrence River region as discussed.

General Description, Distribution, and Members of the Black River Group

The Black River Group was originally defined in the northern Black River Valley near Watertown, northwestern New York State **(figure 5)**. Although first named in this region, exposures of the Black River Valley occur to the southeast along the Black River Valley between the Tug Hill Plateau and the Adirondack Dome and into the Mohawk Valley. They also occur northwestward into adjacent Ontario Canada where equivalent units in the Lake Simcoe region

Figure 5: Type-section localities for Chazy, Black River, and Trenton Group localities for the Lake Ontario to Black River Region of northwestern New York, Southeastern Ontario and southern Quebec. Shown in yellow is the distribution of Ordovician-aged rocks, and in grey are Precambrian-Cambrian aged rocks.

are referred to as the Shadow Lake, Gull River, and Bobcaygeon (in part) formations. To the northeast of the type region in the Lake Champlain region (northeastern New York - eastern Ontario and Quebec), Black River units are also recognized and follow a similar nomenclatural system. However, in the northern Hudson River and easternmost Mohawk valleys, the equivalent units have been referred to using a different nomenclatural system. Historically, this is in part due to substantial thinning, somewhat different lithologic compositions, and less distinctive boundaries between stratigraphic units. Black River Group equivalents are also found in the subsurface to the south and west of the type region and have been correlated over much of eastern North America.

Isopach maps of the Black River (Rickard, 1973) and correlated outcrop sections (Young, 1943 a, b; Cornell, 2001) show these limestones to be thickest along a northeast-southwest directed axis centered in the modern northeastern Lake Ontario Basin. The type region and the Black River Valley lies just to the southeast of the axis, and the exposures in the Kingston, Ontario region lie just to the northwest of the main depocenter. Northeast of the southwest plunging depocentral axis, successively higher Black River formations pinchout against the Precambrian Grenville basement along the Frontenac Arch **(see figure 2)**. To the north and west, Black River formations thin across the Marmora Arch in the vicinity of Peterborough, Ontario before expanding in thickness slightly into the Lake Simcoe region (Liberty, 1969). Likewise to the southeast in the southern Black River Valley and eastern Mohawk Valleys, the Black River is substantially thinner. This is especially the case in areas where the unit onlaps Precambrian basement resulting in the pinchout of much of the basal Pamelia Formation. Likewise, thinning also results where the uppermost units are erosionally truncated at the base of the overlying Trenton Group.

In the Mohawk Valley Black River units are substantially thinner and are completely absent below the Trenton in some areas including at Canajoharie, NY. In similar fashion to the type region, exposures of the Black River equivalents in the Ottawa embayment area show an east-west depocentral trend in the vicinity of the Ottawa-Bonnechere Graben located just east of the Frontenac Arch. Throughout much of the deposition of the Black River, circulation into the Ottawa region was apparently initiated through an easterly connection with the Iapetus Ocean. However, as recognized by Salad Hersi (2000) and Salad Hersi and colleagues (1998), this connection was subsequently closed in the vicinity of the Beauharnois Arch with periodic connections established between the Kingston-Watertown regions across a subdued Frontenac Arch.

Given the distribution and thickness patterns in the vicinity of the type region, it appears that the majority of Black River Group limestones were deposited in the protected shallow epicontinental basin located inboard of the continental margin. However, due to localized subsidence, uplift, and antecedent topographic variations, deposition and facies variations of the Black River were initially restricted to localized sub-basins of the GACB. During deposition of middle Black River formations, this restriction was substantially decreased and units were more widespread and uniform before again experiencing another period of restriction with increased evidence for rejuvenated development of mid-Mohawk Valley topographic features and further activation of the Beauharnois Arch.

As mentioned previously the contact between the Chazy Group and the Black River Group in northeastern New York State and adjacent Quebec represents an unconformity of various extents and duration depending upon location. In the type section of the Black River Group, there is no Chazy below the Black River in northwestern New York State. In most of the region, the Black River is underlain by Lower Ordovician carbonates and sandy dolostones of the Ogdensburg and Theresa Formations respectively. In the southern Black River Valley and adjacent to the Canadian Shield east and north of Kingston, Ontario, Black River units overlap the underlying carbonates and visibly onlap Proterozoic-age (1.0 to 1.2 bya) Grenville gneisses and granites suggesting relict topographic expression during deposition (**figure 6**). This observation is supported by the inclusion of clasts of both Lower Ordovician carbonates as well as Precambrian-derived feldspathic sandstones in basal Black River lithologies in some localities. Thus as exposed in the type region of northern New York and adjacent Ontario, the basal Black River contact represents an unconformity of variable extent, certainly of greater duration than in the Champlain Valley region. Moreover, some evidence suggests that it

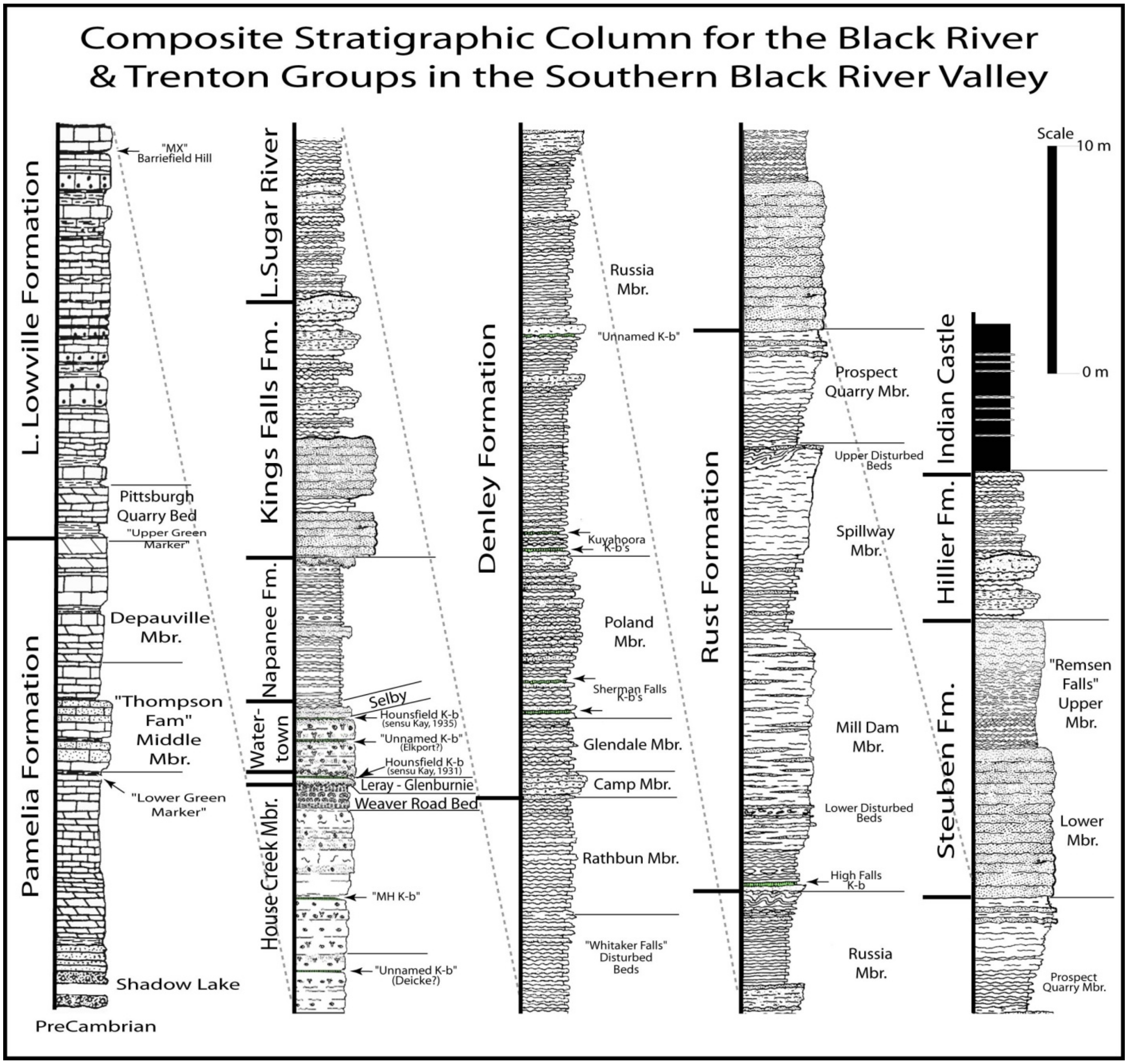


Figure 6: Composite stratigraphic column for the Black River Group in the central Black River Valley (Lewis County) region where it is thinner than in the type area to the north.

represents a slight angular unconformity associated with the Knox unconformity as described in the Mohawk Valley between the Black River and underlying Beekmantown, and Little Falls intervals (Fisher, 1962).

Given the predominance of previous stratigraphic studies, the stratigraphic details of the top contact of the Black River are surprisingly poorly established regionally and are inconsistently applied between the Mohawk and Black River Valleys, the Lake Simcoe region, as well as in areas outside of the type region. Because the position of the Black River-Trenton

Group boundary is still debated in the New York region, it has led to correlation controversies outside of the type region. A major dilemma in this issue has been the paucity of outcrop exposures in the contact boundary interval in the southern Black River to westernmost Mohawk Valleys. Exposures within this critical area are needed in order to trace lateral facies change in the uppermost Black River Group. Recent discovery of a number of cores in the critical region of the eastern Mohawk Valley provides an opportunity for future detailed study. Nonetheless, most stratigraphic studies of the boundary have been completed in the Mohawk Valley region where the Black River Group itself is highly modified and abbreviated by surfaces of non-deposition, extreme facies change, and erosional truncation including karstification. These details are only now being worked out, but historically the position of the Black River/Trenton boundary has been debated for nearly a century (Kay, 1937; Fisher 1962; Titus & Cameron, 1976; Cameron & Mangion, 1977).

Most recently, the boundary has been synonymous with top Turinian or base Rocklandian in the type area (Fisher, 1962). This placement corresponds to a position at the base of the Selby Formation of earliest Rocklandian age and at the top of the Watertown Formation of latest Turinian age. This particular boundary interpretation is drawn on both lithologic and biostratigraphic criteria. In the Black River Valley region, this contact is recognizable as the development of a facies dislocation across a mineralized hardground surface and K-bentonite horizon. When traced into the abbreviated and substantially thinned sections of the Mohawk Valley, the equivalent of the sub-Selby facies dislocation is represented by a very distinctive lithologic boundary. The contact separates off-shore, deeper water facies of the Selby from underlying shallow water facies of the Watertown Formation-equivalent at a pronounced discontinuity (Kay, 1935; 1937). The shallow intertidal equivalent of the Watertown and the overlying deeper-water Selby of the Mohawk Valley shows a pattern similar to the lithostratigraphic and biostratigraphic transition in the Black River Valley. However the lithostratigraphic distinctiveness of this contact is less pronounced in the Black River Valley. In this region, the Selby is substantially thicker and is recognized by a significantly shallower facies that is more similar to the underlying Watertown. Thus recognition of the Watertown-Selby contact is more challenging to place and has not been accepted universally. Consequently, Cameron and Mangion (1977) established the Black River – Trenton boundary position at a higher position. They located the contact at the top of the Selby at the position of the first thin shales and limestones typical of the Trenton Group. In both the Mohawk and Black River Valleys this particular surface is quite pronounced at the first appearance of shales and interbedded calcisiltites, and micritic wackestones.

Yet another contact was proposed by 19th century geologists in the western Mohawk Valley in the vicinity of Middleville. This contact occurs at the top of the "Birdseye Limestone" and base of the overlying Mohawk or Trenton Limestone. For decades this lithologic contact was the important lithologic delineation recognized by geologists. This particular contact was de-emphasized in later studies because the massive-bedded micritic wackestones and calcarenites of the Watertown Limestone, although not equal to the Birdseye, were more similar to it than the overlying thin-bedded shales and limestones of the Trenton. In the Mohawk Valley, where the Watertown equivalent was not previously segregated from the Birdseye, the basal Trenton contact approximated this stratigraphic position. Moreover, without the presence of the Watertown, this assessment suggested that the entire Watertown had been truncated in this region at the base of the Trenton – a position held by Kay. However, through detailed correlations Cornell and colleagues (2005) and Cornell, (2001, 2005) was able to identify a lateral (isochronous) transition in the Watertown from off-shore to more onshore intertidal facies in a southeasterly direction. At the base of the shallow Watertown equivalent facies in the Mohawk Valley, a distinct channeled and karstic surface is documented at Ingham's Mills and other localities (as shown in a number of drill cores from the eastern Mohawk Valley. In the former locality several meters of underlying Birdseye, fenestral micrites and wackestones have been truncated with pronounced erosion.

The same contact in the Black River Valley is also documented by truncation of underlying units, although the amount of truncation is less and the contact is more subtle where the basal Watertown quartz-bearing grainstone is not apparent. Thus although the boundary at the top of the Birdseye has long since been abandoned, there is the possibility (and precedence) for establishing a lithostratigraphic boundary at the base of the Watertown and not at its top. This contact is clearly unconformable with significant erosional truncation along the contact and, although it separates the upper formation of the Black River Group from the lower ones, it brings the contact into alignment with interpretations in the Upper Mississippi Valley, the Jessamine Dome, and the Nashville Dome.

Outside of the New York type region, the complexity of this argument has promulgated an unenthusiastic view of New York nomenclature and there have been attempts to apply, rework, or otherwise abandon the type region classifications for lithostratigraphic and time-rock or stage nomenclature (which means the abandonment of Kay's time-rock terms: Rocklandian, Kirkfieldian, and Shermanian) in lieu of the newer stage called the Chatfieldian as discussed previously. Using this particular term in New York also has an additional complexity in that the base of the Chatfieldian Stage is drawn at the position of the Millbrig K-bentonite, yet this particular volcanic ash is not yet firmly established in New York State (*see* Mitchell et al., 2004;

and Brett et al., 2004 for this particular controversy). Yet regardless of its placement, if the type New York stage classifications can be highlighted and identified on a more regional scale in relationship to the Millbrig using other chronostratigraphic methods such as sequence stratigraphy these stage classifications have precedence over the Chatfieldian.

As a group, the Black River limestones are ~ 60 m thick and have been subdivided on the basis of lithostratigraphy into three formations. In ascending order they are the Pamelia (~40 m), Lowville (~17 m) and the Watertown limestone or Chaumont (~5 m) formations (Fisher, 1977). Many of the early workers, including Kay (1937), and Young (1943 a,b), considered the units of the Black River Group (as well as those in the overlying Trenton) to be both biostratigraphically and lithostratigraphically cohesive. Unfortunately, without a major effort to trace these units from their type sections using substantive internal marker horizons, only minor progress has been made in correlating temporally-constrained intervals across facies transitions (Johnsen, 1971; Walker, 1973; Textoris, 1977). Lacking a well-constrained geochronologic system, the facies distribution of Black River strata came to be viewed as a mosaic of large-scale diachronous facies, with little to no continuity of individual facies over long distances within individual formations. These strata were often considered as time transgressive facies, with component facies grading laterally into each other on a large-scale, time-transgressive Waltherian manner. These models have been tested in recent studies (Cornell, 2001a, and b; Salad Hersi, 2000) on the basis of sequence stratigraphy, K-bentonite and other marker horizon correlations.

Although facies patterns do show evidence for lateral time-equivalent variation, it is possible to differentiate the original lithologic units within biostratigraphically-based successions thus enabling the re-evaluation and recognition of within unit facies transitions and between unit facies transitions. As a result, these studies suggest that isochronous horizons can be identified and often run parallel to formational contacts (at least that coincide with sequence boundaries) without crossing them and in some cases formational and intraformational contacts themselves can even be isochronous surfaces. Clearly, however there is also evidence for diachronous facies patterns (i.e. progradation/retrogradation) within formations and these patterns ultimately help in the recognition of shallowing and deepening trends between successive packages.

Black River Group Formations: Pamelia

In northwestern New York, the Pamelia Formation is the lowest unit of the Upper Ordovician. The Pamelia itself is early Mohawkian in age and rests either on Precambrian Grenville basement rocks or on Cambrian-Lower Ordovician sandstones and carbonates. In either case, the contact at the base of the Pamelia represents a significant unconformity ranging in duration from nearly 600 million years to less than 10 million years. Described from exposures in the Town of Pamelia in Jefferson County (**figure 7**), the Pamelia Formation itself is a significant unit that can be divided into several internal sub-units. These are generally referred to as the lower, middle or "Thompson Farm," and upper "Depauville" members (see **figure 6**). In total, the Pamelia reaches nearly 45 meters in its type region in northwestern New York to eastern Ontario. It thins substantially to the southeast into the southern Black River Valley where the thickness varies but is generally less than 18 meters thick and thins to extinction in some regions south of Boonville, only to reappear in a few locations in the western Mohawk Valley and on the eastern side of the Adirondack Arch.

In terms of composition, the Pamelia is a heterolithic unit containing a wide range of siliciclastic influenced carbonates. Where clearly exposed, the lower member is generally in contact with either the Theresa Formation or the Precambrian basement complex. In either case,

Figure 7: Type area for the Pamelia Formation of the Black River Group. Natural stream outcrops and small abandoned quarries around Pamelia provide access to relatively meager successions, although substantially more strata are exposed in the larger quarry at Lafargeville (Thompson Farm Quarry).

the base is consistently dominated by quartz-rich calcareous sands with fragments of underlying beds imbedded in its matrix. In the upper part of the lower Pamelia, massive planar and crinkly laminated dove-gray micritic limestones appear and are often interbedded with dolomitic caps and green shaly micrites that show an overall shallowing upward pattern to dolostone-dominated beds and mud-cracked argillaceous micrites near the top.

The middle member of the Pamelia formation, informally referred to as the "Thompson Farm Member" for exposures in the Thompson Farm Quarry at Lafargeville, Jefferson County, New York, is distinctive from the lower and upper Pamelia. The middle member contains some quartz-rich beds at its base, but grades upward into medium- to coarse-grained peloidal micrites and bioturbated wackestones and occasional packstones. This unit is clearly a deeper water facies and is recognized through the appearance of bryozoans, large-cephalopods, brachiopods, a variety of bivalves, and some crinoid ossicles and occasionally an entire crinoid itself.

Moreover, in places this unit may contain *Tetradium* coral colony thickets and an occasional stromatoporoid. In the southern Black River Valley the middle Pamelia also contains intervals of ooid grainstones.

The uppermost member of the Pamelia, named for the Depauville Waterlime of early geologists, was described as gradational out of the middle member into the overlying Lowville Formation. As defined the base of this unit occurs at the first massive earthy-weathering dolomitic limestone and extends up through the uppermost dolomitic limestone (Cushing, 1911, Young, 1943). At Depauville, NY, the thickness of the unit as originally defined is difficult to establish owing to incomplete and covered exposures, but is measured to be at least eight meters thick in the outcrop just north of the village on NY Rte. 12, and in the Thompson Farm Quarry to the east of Depauville, it is about the same. Although this expanded Depauville contains all of the predominantly dolomitic beds, there are significant patterns (shallowing-upward followed by deepening-upward), lithologic breaks, and important stratigraphic surfaces within the Depauville that warrant further refinement. Based on several key contacts and correlative marker horizons, the Depauville is restricted to the most substantial massive-bedded dolostone interval with extensive vugs below the Pittsburgh Quarry Bed of Conkin (1991). The vugs are typically filled with celestite (strontium sulfate) nodules and/or other evaporitic pseudomorphs. In some cases, although no minerals remain in weathered sections, it is not uncommon in bedding cross-sections to see evidence for moldic preservation of cubic or bladed evaporite textures similar to halite or gypsum impressions.

In similar fashion as the top of the lower Pamelia, the top of the Depauville member is demarcated at the base of a distinctive recessive-weathering, green dolomitic silty mudstone and micrite unit containing small well-rounded, poly-crystalline quartz grains and sometimes a reddish or magenta shale sitting on top of the subjacent dolomitic limestone. This bed has been referred to as the "Upper Green Marker" in the Lake Simcoe area or the "Pittsburgh Quarry Bed" in southeastern Ontario by Conkin (1991). In the Black River Valley close to the Adirondack Arch, this interval can also show feldspathic sandstones derived from local Precambrian exposures. The contact contains some evidence for truncation at the base of the overlying Lowville Formation.

<u>Black River Group Formations: Lowville</u>

General Comments

The Lowville Formation was named for exposures of the "birdseye limestone" in the vicinity of Lowville, Lewis County, New York (Clarke and Schuchert, 1899) (**figure 8**).

Figure 8: Type locality for the Lowville Formation on Mill Creek Lowville, New York. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the west.

Cushing et al. (1910), included two main units in the Lowville. The lower Lowville was the "birdseye limestone" of the early workers on the basis of its most characteristic lithology –

fenestral micrites. The upper Lowville, later renamed the House Creek member, was substantially coarser and more fossiliferous. At present the lower, typical “birdseye micrite” remains unnamed and will be referred to as the lower Lowville member. In southwestern Ontario, the lower Lowville is equivalent to much of Liberty’s (1969) middle Gull River Formation (his B1 and B2 submembers, but not B3; Cornell, 2001).

The Lowville Formation has a very widespread distribution and nearly identical facies can be observed from the eastern Mohawk Valley of central New York State through the westernmost part of the study area. Its lithologies are very distinctive and in many cases other units have been given the Lowville name even in cases where they are clearly not age equivalent to the type Lowville Formation. In the Black River Valley outcrop belt it typically ranges between 10 and 14 m (30 and 45’). To the southeast at Ingham’s Mills only about 4-5 m is exposed. In the eastern Mohawk Valley region, the Lowville has been demonstrated to thin very drastically and it has been truncated completely beneath a complex unconformity in the Canajoharie region. Beyond the Mohawk Valley the Lowville reappears in the area around Amsterdam, New York. In Ontario, the equivalent Gull River B1-B2 beds bracketed by the MX and MH K-bentonites, also thin toward the west of Lake Simcoe. In this region the thickness is ~10.5 m.

Lower Member of the Lowville Formation

The Lowville Formation contact with the Pamelia Formation was originally drawn at the top of the highest development of dolostone (Cushing et al. 1910). This description can be problematic depending on locality. In some cases, dolostones in the Pamelia, including in the Pamelia type region, appear to be primary dolomites and follow bedding patterns. Others appear to have been dolomitized by secondary fluid transport especially in the vicinity of known fault zones, as well as along the Precambrian contact. Therefore the position of dolostones can be variable throughout the area. However, Young (1943) provided a more concrete position for the contact of the Lowville with the underlying Pamelia Formation. In the region of Roaring Brook, in the central Black River Valley, he noted that the basal Lowville to contain a conglomerate within lenses of soft greenish mudstone to shale and described the contact as disconformable. This horizon is now used as a stratigraphic marker for correlation across the study region and is coincident with the "upper green marker" bed discussed previously (**see figure 6**). The basal contact is sharp and well-defined by the change from massive buff dolostones, often with a red stained pyritic top, to a very sandy quartz rich green mudstone. In some locations, this siliciclastic dominated mudstone contains variable sized clasts of Pamelia carbonates and can have well-rounded quartz, feldspar, and granitic pebbles derived from the Precambrian. The top contact of the lower Lowville member (B1-B2 sub-member of the Gull River of Liberty, 1969) with the House Creek Member (Moore Hill/B3 sub-member of Liberty, 1969) occurs at a distinctive horizon that recognized over the region. The contact is defined by the sharp change to fine-grained, deeper water, wackestones and packstones of the House Creek/Moore Hill (see below).

The lower member, in the type region consists of about 13 m of sparsely fossiliferous, fenestral dove gray micrites and minor shales. Minor interbedded buff-weathering, mud-cracked, dolomitic limestones generally cap thin shallowing upward cycles especially in the lower half. Occasional ooids, ostracods, crinoid plates, trilobite fragments and bivalves make up the allochems and become more prevalent in the upper part. The spar fractions of most samples are present in either of two forms – void space birdseye fillings or replaced calcite spars. The term "birdseye" comes from the calcite spar filled vugs and burrows in the fine cryptocrystalline matrix. These features can be cross sections of *Phytopsis tubulosa* burrows, or are void space fenestral fillings. The later are generally considered birdseye structures (fenestrae) sensu stricto by modern carbonate sedimentologists. Using the Folk classification system, the Lowville would be termed interbedded pelmicrites, pelsparites, and biopelmicrites. Also found in most outcrop areas are beds of intramicrite, biopelsparite, and *Tetradium* biolithites especially near the top of the lower Lowville member (Textoris, 1968). Occasionally in the lower Lowville thin beds of dolomitized micrite occur. Thus defined, the carbonates of this interval, and indeed throughout much of the Black River are dominated by peloidal micrites whether in New York or in the Lake Simcoe District of Ontario (Grimwood et al., 1999). In addition to this lithologic character, this interval is also known for well developed, parallel laminations often in thin-thick patterns, rare wave ripples, well-developed mud cracked horizons especially at the tops of cycles, and normal graded bedding. In addition to *Tetradium*, other faunal components are minimal but include small gastropods, ostracods, fragmented bryozoan colonies, and a few *Bathyurus* sp. trilobites all of which suggest a fairly normal marine environment of deposition.

House Creek Member of the Lowville Formation

The upper Lowville Formation has long been recognized as distinctive from the lower member. Ruedemann & Kemp (1910) named the upper Lowville the Leray Member for sections in a quarry near Leray Street in northern New York (**figure 9**). The Leray contains well- bedded chert horizons near its top below the massive bedded Watertown Limestone. Outside this immediate vicinity, the Leray was difficult to separate from the Watertown especially where the intervening Glenburnie shale was not prominent or well-weathered. Thus, Kay (1960) proposed the Chaumont Formation to encapsulate the two intervals. However, many workers have not recognized this unit, and have favored the definition of Cushing (1910) that placed the upper

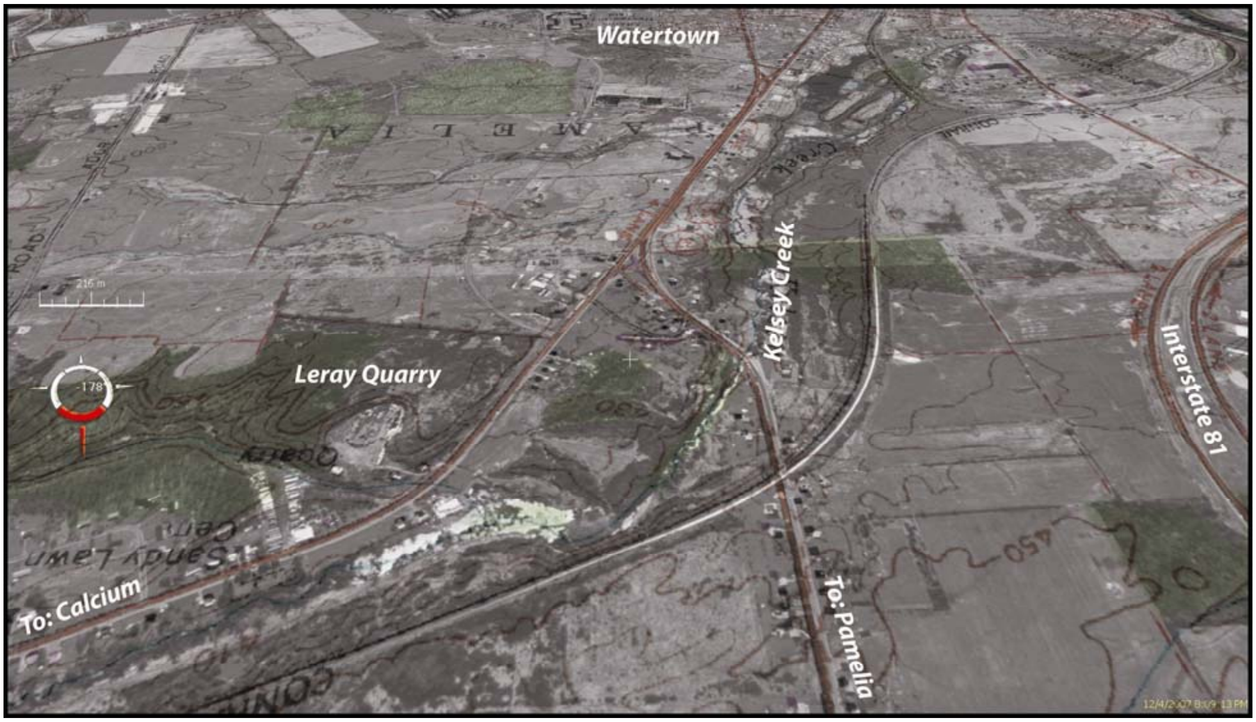


Figure 9: Type locality for the Leray Member of the Chaumont Formation on the east side of NY Route 11 just north of its junction with NY Route 3 and approximately 1.2 km east of I-81. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the west.

limit of the Lowville at the base of the massive-bedded Watertown Limestone. In outcrop regions near Ottawa, the term Leray is still utilized for the entire interval through the base of the Rockland, up to and including the Watertown Formation of New York. Thus, the upper Lowville was renamed (Walker, 1973) the House Creek Limestone Member and distinguished the unit on the basis of the dominant lithology of the interval which was the light to medium-gray, medium-bedded, bioturbated wackestones that often contain densely packed *Tetradium* coral biostromes. The term Leray has generally been abandoned in New York; however, near Watertown the Leray of early workers forms the uppermost interval of Walker's House Creek. It is distinctly coarser-grained than the typical Lowville, although it appears to be substantially-thinned and is eventually truncated in the central and southern Black River Valley at the base of the Watertown.

As currently recognized, the upper House Creek member is comprised of medium gray to brown, burrow mottled, locally cherty wackestone to packstone. The House Creek is about 4.5 m (13-15') and the equivalent Moore Hill Member of the Gull River Formation in Ontario, is about 5 m (12-15'), but can range up to ~6 m locally. In the southeastern most exposures of the Mohawk Valley the thickness of this unit thins and is about 2.5 m (8') at Ingham's Mills where most of the upper portion is erosionally truncated at the base of the Watertown Formation.

The House Creek Member contains evidence for a more off-shore, yet still shallow, depositional environment. The unit commonly contains abundant *Tetradium*, large tabulate corals including *Foerstephyllyum* sp., and *Lambeophyllum* and, stromatoporoids, as well as associations of gastropods, bivalves, and a few brachiopods. The characteristic bioturbated biostromal *Tetradium* beds can be recognized over a very widespread area and together with the MH K-Bentonite demonstrate that this particular interval is equivalent to the Moore Hill (Okulitch, 1939) of Lake Simcoe. As a unit then the House Creek/Moore Hill can be described as medium-grained, pale brown to dark gray fossiliferous limestones that are distinctively coarser-grained than underlying units.

The upper contact of the cyclically bedded House Creek/Moore Hill interval is marked by the transition into, the informal Weaver Road beds. This transition is generally drawn at the base of a 0.75-2 m (2-6') shaly zone. In the eastern portion of the study region (Kingston through northern New York State), the upper Moore Hill likewise grades into the platy dark shales and interbedded domal stromatolitic wackestones of the Weaver Road beds. In the Lake Simcoe district, this same horizon is recognized by a thin shaly nodular horizon that contains oncolites at the base of the Coboconk Limestone. This horizon is developed over the typical massive bioturbated packstones and wackestones of the underlying Moore Hill Member. In the region between Coboconk and Marmora, this same contact is developed again from the transition from the massive bioturbated beds of the underlying Moore Hill into thinner bedded wackestones and interbedded shales which often show well developed mudcracks.

Weaver Road Beds

A particularly important marker horizon in the uppermost House Creek Member comprises a thin interval of mud-cracked, gray-yellow weathering shales (similar to kukersites?), platy micrites and domal stromatolites. This unit was designated the "Weaver Road Beds" by Cornell (2001). This interval is recognized over much of the northern Black River Valley as well as in adjacent Ontario. It is very well exposed in the Black River Gorge at Glenn Park (when water levels are low). The interbedded shales and domal stromatolites of this unit can be observed in relation to the underlying bioturbated interval of the House Creek and the overlying Leray and Watertown. In a drill core from the region south of Bath, Ontario, Noor (1989) recognized this same interval and referred to it as the Bath sub-member of the Bobcaygeon Formation. In most of the northern Black River Valley successions of New York, this unit is present although it loses the characteristic domal stromatolites which grade into less prominent LLH style or laterally linked hemispheroids in the southern Black River Valley. It does, however, become truncated below the overlying unit southward into the Mohawk Valley. Overall, the biofacies and the lithologic character of the House Creek indicate a shallowing-upward pattern.

In the reference locality, the Weaver Road is just less than two meters and maintains this thickness through the Kingston, Ontario region (Cornell, 2001). To the south of the type region the shaly Weaver Road interval thins substantially and loses the domal stromatolitic component. The dark shales grade into thin bedded, laminated lutites and shales (similar to the Lowville birdseye limestones) near Lowville, but are distinct in that they lie above the House Creek rather than below it. At Boonville, the Weaver Road interval is no longer present and is interpreted to be erosionally truncated (Cornell, 2001).

Westward into Ontario, the accessory grey-black shales thicken towards Napanee, Ontario, where they become a more dominant component and have a maximum thickness of just over 2 m. Farther west, the Weaver Road thins substantially and is found at Marmora as a thin, shaly nodular unit. It is interbedded with thin calcisiltites that are typically desiccation cracked and vertically burrowed and are ~ 0.5 to 0.6 m thick. Near Coboconk, Ontario, the Weaver Road becomes an argillaceous rubbly-weathering carbonate (~1.1 m thick) just below the base of the Coboconk. Farther west, in the region of Brechin quarry, the Weaver Road has not been explicitly documented as a shaly facies. However, north of Lake Simcoe (Medonte, Ontario), Kay (1931) recognized this interval to contain 3’8” of “interbedded, fine textured, very light gray limestones and blue-gray, laminated, papery, tough shales.” This description is very similar to that observed in eastern Ontario in both the Glenburnie road cut and quarry, as well as in the Highway 401 road cut at Napanee. Its occurrence indicates that this restricted horizon is fairly continuous and similar in lithology across the region, although it is not as well-developed or maybe truncated at Brechin on the western side of Lake Simcoe.

The upper contact of the Weaver Road is often an irregular, sharp horizon. In the Lake Simcoe region, the contact with the Coboconk is fairly sharp. At several localities including Brechin Quarry it is delineated by the presence of a well developed hardground with *Solenopora* overlain by a much coarser (packstone to grainstone lithology) often with numerous stromatoporoids. In New York, the coarse-grained interval overlying the Weaver Road is the Leray Limestone of earlier workers.

Black River Group Formations: Chaumont

The Chaumont Formation (of Kay, 1929) or the expanded Watertown (of Walker, 1973, Fisher, 1977), like other Black River units, is composed of a series of internally differentiated sub-units characterized on the basis of both litho- and biofacies. As recognized by Cornell (2001) the Chaumont Formation is composed of three members based on a reconciliation of historic assessments and improved correlations.

In the Glenburnie and Kingston, Ontario region the basal contact of the Chaumont is sharp and overlain by a ~ 0.5 m bed of strongly cross-bedded grainstone with a minor quartz sand component. This same scenario repeats in northern New York State, where the lower Coboconk quartzose grainstone appears in the Black River Gorge at Glenn Park and in the nearby county Rte. 54 road cut at Brownville where it forms the bottom of a couplet of distinctive beds that form the base of the Chaumont Formation. In New York, the lower Chaumont is a 0.5-l.5 m interval of crinoid/gastropod quartz-rich grainstone and superjacent welded coral and brachiopod rich packstone. These rest sharply and disconformably on the Weaver Road interval. Although originally described as a chert-rich unit in the type area, this interval is more typically characterized by the presence of a quartz-rich basal bed with disseminated detrital quartz grains in the overlying few centimeters.

Coboconk Member

In the Lake Simcoe district of Ontario, the Moore Hill/House Creek interval is overlain by the "Coboconk Member" of the lower Bobcaygeon Formation. The Coboconk Member (proposed by W.A. Johnston; 1910) is defined as "dark blue to gray nodular and chert[y] limestones" and was placed in the Black River Group by Okulitch (1939). However, the Coboconk was situated across the Black River –Trenton transition and a specific contact was not formally recognized, although most workers considered the Coboconk to be equivalent to the Watertown and the lower Trenton. Subsequently all units were placed in the Simcoe Group for mapping purposes owing to the difficulty in placing the Black River/Trenton boundary (Caley and Liberty; 1967). Nonetheless Okulitch (1939) had recognized that the Coboconk was internally divided by a distinct disconformity especially prominent in the type locality and suggested that the name could be restricted to the basal portion of the unit. Following his suggestion, Cornell (2001) proposed that his beds 14 and 15 ~ 2.3 m (7') thick retain the name Coboconk. Thus the ~4.3 m thick uppermost bed (bed 16) was excluded from the Coboconk member. The distinctive sharp, contact noted by Okulitch sets massive bed 16 apart especially in locations where bed 16 shows evidence of erosion and contains lithoclasts of subjacent lithologies. Liberty (1969) also noted the distinct intraclastic lithology and used it to separate his C-1 sub-member from the overlying C-2 sub-member. Bed 16/C-2 is thus assigned to the Watertown Member, and the contact between beds 15 and 16 (C1-C2 respectively) is interpreted as an erosion surface, and an important sequence boundary.

This contact is traceable eastward in Ontario. In the Napanee to Kingston region, the lower Coboconk beds are separated from the upper beds by a new unit below the level of the unconformity. This new unit is dominantly a shale unit that contains thin interbeds of micrites and is referred to as the Glenburnie Shale. This latter interval is considered the medial member of the Chaumont Formation (Kay, 1929; 1931; Cornell, 2001). As defined, the basal member of the Chaumont Formation is referred to as the Coboconk Member and is equivalent to the Leray of northern New York. Although the cherty nature of this unit is variable and not consistent in all localities, this unit has unique taphonomic characteristics that enable its distinction. Fossil materials in this unit, including gastropods, brachiopods, corals, and bivalve fragments, are often preserved by early silicification. In weathered outcrops they often weather in relief to the surrounding matrix making the coarser grained facies even more pronounced and set off compared to the underlying units. South of the type region, the basal Watertown is significantly less obvious without the substantial quartz component and appears to transition to a series of thinner-bedded wackestone beds before pinching out between the southernmost Black River region and the Middleville, NY in the West Canada Valley.

The Coboconk (sensu stricto), as recognized by Cornell (2001) has a cumulative thickness at the type section of slightly more than 2.5 m (~7'). As the Coboconk is traced west of Lake Simcoe, it is recorded by a thin interval of, thin, interbedded fine-grainstones and shaly calcisiltites suggesting deeper facies and at Brechin Quarry the interval is about one meter (3.25') thick. Eastward of Coboconk, toward Marmora, Ontario, the Coboconk thickens and picks up signatures of slightly shallower facies, and the grainstones show an increased occurrence of corals and stromatoporoids. The crinoidal grainstones and stromatoporoid beds developed in the Coboconk region transition into poorly sorted ooid grainstones interbedded with colonies of *Tetradium*, stromatoporoids, and favositid tabulate corals. In Marmora, the Coboconk is 3.5 meters (10') thick. Toward Napanee, the Coboconk thins and is overlain by the Glenburnie shale. At Glenburnie, Ontario, the Coboconk is a thin (less than 0.5 m thick) welded bed with a basal coarse grainstone with herringbone stratification and weakly developed climbing ripple marks. Its cap is a condensed wacke- to packstone lithology containing a number of taxa including brachiopods, corals, and bivalves. Many of which are shared with the overlying Watertown and the Trenton Group rather than the underlying Black River. In this region, the couplet of beds is also overlain by the Glenburnie.

Glenburnie Member

Immediately above the Leray/Coboconk (sensu Cornell, 2001) is a 0.5 to 1 meter thick shale to shaly nodular limestone interval with a distinctive fauna (**figure 10**). Although not well-developed in most localities southeast of the village of Brownville, northwestern New York, it is recognized in sections as far south as Boonville, New York where it separates the massive 7' tier

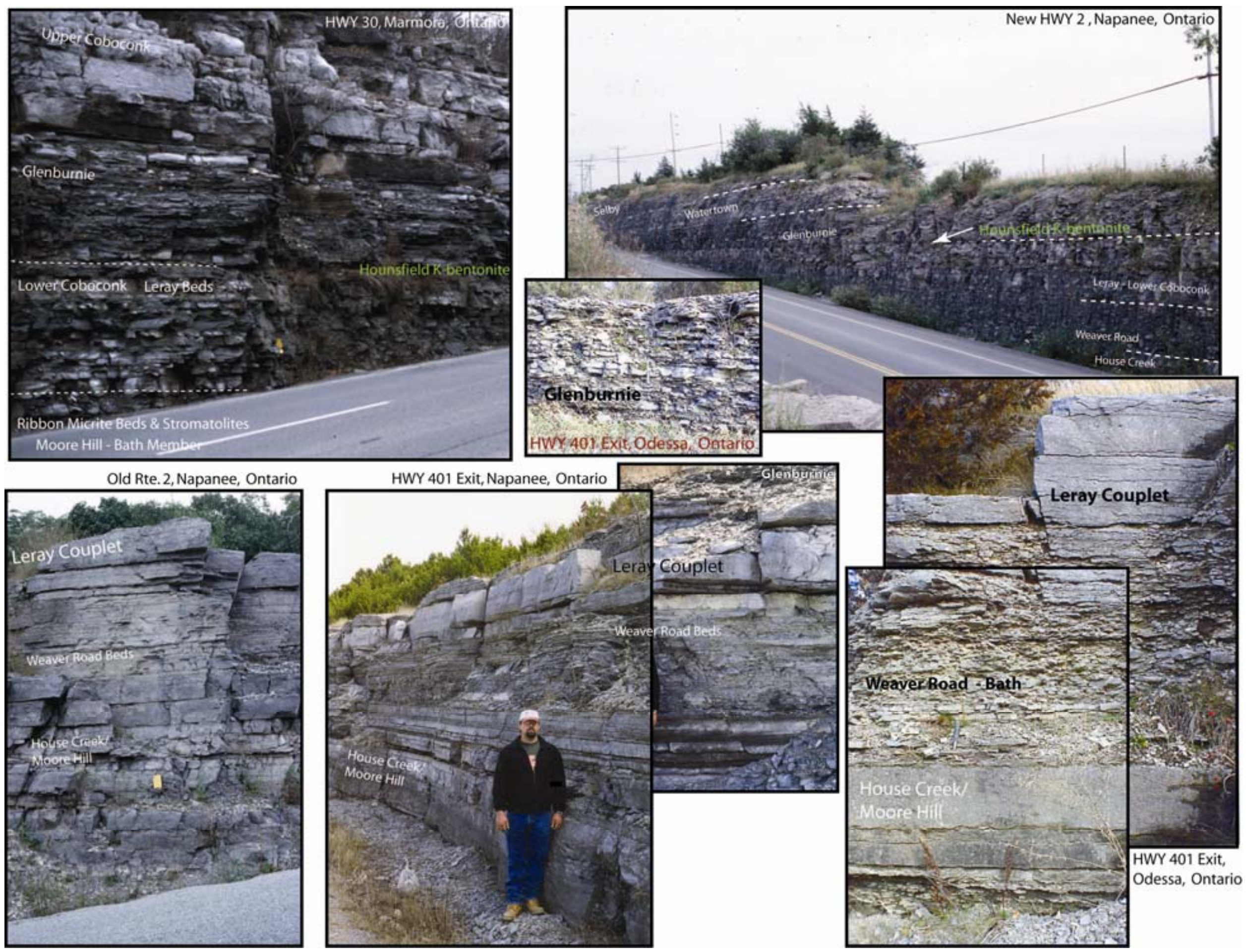


Figure 10: Outcrop exposures of the Weaver Road-Leray-Glenburnie-Watertown succession at Napanee, Odessa, and Marmora, Ontario.

of the Watertown Limestone from underlying units. The shaly-nodular facies of the middle Chaumont Formation does become more pronounced into Ontario where this unit has been named the Glenburnie Shale.

At Glenburnie about 60 centimeters (~2') of dark grey-black, organic-rich shales sit on the grainstone-wackestone couplet, in the road cut along Unity Road, just west of its intersection with Battersea Road and in the quarry located to the southwest (on McKendry Road). Overlying the shales is an intraclastic, conglomeratic grainstone bed. The base of the shale contains the Hounsfield K-bentonite (sensu Kay, 1929) and was used for correlation in Ontario and even with the Decorah Shales of the Upper Mississippi Valley. The dark Glenburnie Shale and its typical brachiopod-bryozoan rich fauna are recognized at Marmora, and again in Brechin Quarry where it is slightly thicker (~1m). To the southwest of Kingston, Noor (1989) identified the same interval from core and included this shaly interval along with the lower shaly interval (Weaver Road herein) in the "Bath Member" of the Bobcaygeon Formation. Southward of its type region near the north shore of Lake Ontario, the Glenburnie and subjacent interval appear to be thicker (Noor, 1989) than in outcrop in the vicinity of the Kingston Trough.

The Glenburnie was not initially recognized by Kay in New York. Nonetheless, Ulrich (1910) had reported a dark shaly interval with well-developed bryozoan faunas at a locality near Three Mile Bay, New York. These were similar to those of the Decorah Formation of Iowa, Minnesota, and Wisconsin. Subsequently, although more nodular and less shaly, this same interval has been traced to Brownville, NY where it is very thin (<.3m) nodular limestone overlying a recessive weathering notch containing a K-bentonite (**figure 11**). The bentonite has not been recognized at Lowville, New York, but it does occur in Boonville in the southern Black River Valley. At Brownville, and in some locations in the Barrett Paving quarry at Boonville, the K-bentonite can be observed weathering out of fresh outcrop surfaces. These bedding planes are occasionally opened by quarry operations and up to 4-5 cm of the K-bentonite can be present in irregular depressions on the surface. At Brechin, Ontario, the approximate position of the

Figure 11: Outcrop photographs of the House Creek-Leray-Glenburnie-Watertown Interval of New York from the Brownville area in the northern Black River Valley to Boonville in the southern Black River Valley.

Glenburnie is marked by fossiliferous shales interbedded with thin wackestone – packstones beds suggestive of Glenburnie lithologies and faunas. The Hounsfield K-bentonite is found near the base of this interval. In some instances, (i.e. Miller Paving Quarry at Dalrymple, Ontario) the K-bentonite is discontinuous for it can be observed in the eastern quarry wall, while it is absent in the adjacent (northern) and western walls.

In many New York outcrops, without a distinctive shaly Glenburnie interval, the Chaumont Formation appears to be gradational from the grainstone/wackestone couplet (of the Coboconk/Leray Member) upward into the massive "upper tier" of the Watertown Member. However regional study shows that this interval actually is separated from the Watertown by a significant unconformity which can be elucidated by the presence of the intraclastic grainstone bed – often with quartz granules and occasional evidence for solution pitting and channeling of the underlying beds.

The upper contact of the Glenburnie with the overlying Watertown is distinctive. At the type section, the contact is knife sharp and when traced laterally along the outcrop it shows local angular truncation (~10 to 20 centimeters) of the Glenburnie. Due to the recessive weathering of the shaly-nodular Glenburnie interval, the overlying Watertown weathers out in relief. This horizon is well developed in the Highway 401 cuts at Kingston. In sections near the village of Three Mile Bay, New York, outcrops are well weathered and overgrown in old abandoned quarries on both ends of the village. Nonetheless, the contact is sharp and it appears to have truncated some of the underlying Glenburnie as in this case there are occasional thin clay-rich clasts weathering out of the basal Watertown. This contact eventually cuts out the Glenburnie, subjacent Coboconk, and even underlying House Creek lithologies. In the vicinity of Middleville, New York, and again at East Canada Creek there is evidence of channeling or truncation of underlying units.

Where it is best developed, the Glenburnie Member is dominantly dark gray papery shale with interbedded calcisiltites and minor bioclastic wackestones. Carbonate beds of the Glenburnie are bioturbated and contain abundant fossils in the calcisiltites and wackestones. It shares many faunal similarities with the underlying Coboconk/Leray and the overlying

Watertown. In Ontario, it contains abundant bryozoans (*Escharopora*, *Hemiphragma*, *Pachydictya*, *Rhinidictya*, etc.), several brachiopod taxa (*Strophomena* sp., and *Rhynchotrema*), a number of corals (*Streptelasma profundum*, *Columnaria*, and *Tetradium fibratum*), several bivalves, and only two nautiloid taxa (*Orthoceras recticameratum* – a form common in the Lowville; and *Spyroceras bilineatum* a form known from the Watertown and Trenton of New York and Tennessee). Several species of trilobites are present and are a number of ostracods (Kay, 1931; listed 12 different species from five genera). Overall, compared to underlying units, the Glenburnie appears to be transitional between that of the Black River Group and with that of the Trenton Group and is a somewhat deeper facies than underlying units.

As mentioned, the Glenburnie member was also described (*sensu* Kay, 1931) to contain the Hounsfield K-bentonite. Subsequently Kay (1935) relocated the Hounsfield K-bentonite to a higher stratigraphic position at the contact with the Watertown and Selby. Nonetheless, there is a K-bentonite in the Glenburnie at the level described by Kay (sensu 1931) and here is referred to as the Hounsfield sensu Kay, 1931). This same horizon has been sampled for chemical fingerprinting and has shown similarities to the Millbrig (see Mitchell et al., 2004).

Watertown Member

In the type area, a five meter thick interval of massive, ledge-forming condensed packstone to micritic grainstone beds occur above the Glenburnie/Leray interval and contains intervals of dark grey to black bedded cherts. This unit, with a distinctive coral, algae, brachiopod, and cephalopod fauna, was termed the Watertown Limestone by Cushing et al., (1910). The Watertown has generally been assigned to the Black River Group as it appears lithologically more similar (more light-medium-grey carbonate, more massive bedding, and less shale) to the underlying strata than the overlying Trenton. Conkin (1991), however, has argued that this unit should be assigned to the Trenton Group based on faunal and lithologic evidence and to equilibrate the type section with equivalent sections in central Kentucky.

The Watertown Limestone was originally defined from exposures in the Black River Valley near Watertown, New York. The massive limestone unit making up the main waterfalls at Lowville, and Watertown, was called the Black River Limestone (Hall, 1847). In his usage, it was originally equivalent to the "Seven-foot tier" of the quarries he visited (**see figure 11**). Since Hall, most workers used the term Black River more broadly for the carbonate rocks in the region. Cushing and colleagues (1910) proposed the Watertown Formation to encompass the 7' tier as well as (2.5-3.0 m) of underlying dark limestones, which included the Glenburnie Member and the Leray. Cornell (2001) redefined and restricted the Watertown to represent the massive bedded cephalopod rich limestone above the Glenburnie. Thus Watertown is only to be used for the equivalent of Okulitch's (1939) bed 16 and the approximate interval of Hall's "7' tier" of the southern Black River Valley.

The Watertown limestone and its equivalents are widespread in New York and Ontario. The Watertown reaches a maximum thickness of about five meters (15') in northern Jefferson County in the vicinity of Chaumont. It is more commonly about three meters (10') thick, and thins to two-three meters (7-8') in the southern Black River Valley where it received its fame by quarry workers as the "7' tier". By the position of Ingham's Mills in the East Canada Creek Valley, the Watertown has thinned to less than two meters. In Ontario, the Watertown becomes more coarse-grained and less bioturbated so that in the Kingston to Napanee area most exposures show more grainstone rich intervals, with some oolitic horizons, especially in the Napanee region. In the latter region, these coarse facies have often been considered equivalents of the Coboconk but represent only the upper portion of the Coboconk of earliest workers.

Nonetheless, black nodular to bedded chert horizons are still found in the Watertown and its equivalent units toward the west, and can be used as general marker horizons.

When traced into the southern Black River and Mohawk River Valleys, the Watertown becomes substantially thinner, more fine-grained, and even shows a transition from offshore deeper-water facies into intertidal facies as displayed at Ingham Mills and a few other western Mohawk Valley localities. Lithologically, these Watertown equivalent limestones are very similar in appearance to the lower Lowville Formation and have been referred to as such in many localities. Yet by tracing out individual packages and marker horizons it is possible to differentiate Watertown equivalent shallow water facies. This lateral facies change is identifiable and follows the diachronous concepts of Fisher (1962) and Walker (1973) but only within high-order cycles, and not on the formation-scale as previously proposed. In this region, as in the type area, the contact at the base of this unit is recognized by noticeable incision and channelized truncation of underlying beds followed by a series of intraclastic conglomeratic beds. Collectively these observations indicate that there may be the development of another interval of erosion that is more pronounced in the southern Mohawk Valley.

The contact of the Watertown Formation with the overlying Selby Formation appears to be gradational in northern New York and Ontario. It only appears sharp in the southeastern portion of the study area (Watertown to Lowville) because of the amalgamated nature of the massive “7 foot tier” of the Watertown Formation. In the region farther to the south (Boonville), and in the Mohawk Valley region, the contact with the Selby has previously been described as being unconformable, but it appears to be conformable and similar to minor flooding surfaces seen elsewhere in the lower succession.

Biofacies of the Watertown Formation are distinctly more diverse than any of the underlying units. The Watertown retains a variety of coral species, as well as a number of taxa from the underlying House Creek and Coboconk, yet a variety of additional taxa are present. A number of bryozoans, cephalopods and various algae are introduced rather abruptly in the Watertown limestone (some new taxa first appear in the Glenburnie). According to Cushing and colleagues (1910) the Watertown "is essentially a cephalopod facies." The nautiloid *Gonioceras anceps*, and several species of *Endoceras* can be found in just about every outcrop where there is a flat bedding plane exposed. Well over seventy cephalopods occur in one instance alone on a bedding plane approximately twenty meters by fifteen meters in width. This diverse cephalopod fauna and widespread cephalopod pavements may coincide with the large-scale extinction of these mollusks as described by Whalman (1992). Based on its faunal similarity to the overlying Trenton – Kay (1929) favored the idea that the Watertown be allied with the Trenton Group although on a lithologic basis it lacks the relatively high proportion of siliciclastic and carbonaceous interbeds that define the Trenton Group.

General Description, Distribution, and Members of the Trenton Group

The Trenton Group and its equivalents were deposited during the largest overall sea-level rise events in the Upper Ordovician (Ross & Ross, 1992, 1995). During this late Mohawkian time, much of the Laurentian craton was submerged including areas well up into the Canadian Shield – areas that throughout the Ashbyan and Turinian, were elevated and exposed to weathering **(figure 12)**. Kay (1937) and Sharma and colleagues (2003) have suggested that, at the very least, portions of the Canadian Shield region to the north not only saw the effects of Late Ordovician sea-level change, but that these areas were also connected hydrodynamically to the Taconic Foreland Basin. Moreover, it is now also noted that even far removed areas such as the Timiskaming region (Timiskaming Strait of Kay, or Timiskaming Graben of Sharma et al) may also have seen the tectonic and sedimentologic impact of Taconian tectonics. Nonetheless, understanding the original distribution of Trenton Limestone deposits is a challenge due to post-Ordovician tectonic activity and subsequent erosion. However small outcrop outliers i.e. The Wells Outlier in the southern Adirondacks (**see figure 3**) and the Timiskaming outlier in

Figure 12: Paleogeographic interpretation for the type region of the Chazy, Black River, and Trenton groups during the Late Ordovician and during major onset of Taconic orogenesis –during the greatest extent of Trenton deposition. Key regional features are located as recognized from sedimentologic and stratigraphic evidence.

Ontario help provide important clues to the original paleogeographic distribution of these rocks.

Farther to the west, it is known that limestone deposition with minor siliciclastic contribution extended through the Michigan Basin (where major eastward tilting was shown to have occurred during the late Mohawkian; Coakley & Gurnis, 1995), across the Upper Mississippi River Valley, and perhaps even across the Transcontinental Arch (Fanton & Holmden, 2007). Southwest of the type Trenton region (**see figure 1**), Trenton equivalent limestones become mixed with substantially more siliciclastics. Along the northwestern margin of the Champlain Trough on the Beauharnois Arch, and along its southward extension, the Adirondack Arch (see Figure 18), recognizable type-Trenton limestones are prevalent only during the Rocklandian. Subsequently, carbonate contribution is substantially reduced and the siliciclastic components are increased (shales, siltstones, and even sandstones) so that the lateral equivalent of the mid-upper Trenton is referred to as the Martinsburg flysch. These sediments seem to emanate from regions to the southeast of the carbonate platform. To the west of the Adirondack Arch, within the Olin Trough (northeastward extension of the Rome Trough) Trenton limestones interfinger with dark shales and show significant evidence for southwestward transport (from shallower areas to the north). It appears that in central Pennsylvania, the Adirondack Arch and Olin Basin represented significant paleotopographic features that were important in influencing the deposition of Trenton limestones. Moreover, the impact of Taconian tectonism on the Laurentian craton and the GACB was fully realized across the craton by the time the Upper Trenton limestones were deposited.

The Trenton Group is defined on the basis of highly fossiliferous interbedded coarse to moderately coarse-grained limestones, less fossiliferous micritic or calcilutitic limestones, and interbedded calcareous shales throughout its depositional area. Its formations and members are recognized on the basis of variations in the bedding style and relative abundance of each of these different lithologies (Brett and Baird, 2002). Collectively, the facies of the Trenton Group represent a range of depositional environments: from relatively shallow water high-energy environments to deeper ramp, turbidite-influenced facies (Cisne, et al., 1982; Lehmann et al, 1994). Given the sedimentologic distinctiveness of the group and its stratigraphic position relative to the shallow water facies of underlying Black River Group and the overlying black shale-dominated facies of the Utica Group, it is relatively easy to identify in outcrop and in core. It is also fairly easily distinguished in well-logs so that the interval is fairly well established in many outcrops and in the subsurface. Nonetheless, the nature of its lower and upper contacts, have been the center of early and ongoing debates (as discussed earlier) and are still problematic.

The lower contact of the Trenton Group with the underlying Black River Group has been discussed previously. The upper contact of the Trenton Group is perhaps best considered by Brett & Baird (2002). In the type sections at Trenton Falls, the uppermost contact of the Trenton is not well exposed; however, sections north and east of Trenton Falls have been used to fill in the contact information. In this region, the uppermost Trenton is easily recognizable as a massive cross-bedded coarse crinoidal limestone. The top contact of this unit with the overlying Indian Castle Shale is demarcated as an erosional disconformity by Baird and Brett (2002). The presence of the phosphatic lag at the top of the Steuben, the Honey Hill bed, and in lateral equivalents down ramp to the east, provide evidence for submarine erosion that represents a significant gap in sedimentation before the Trenton shelf became completely smothered in the type region by dark anoxic mud of the Indian Castle Shale.

To the east of the type section, the middle to upper Trenton Group (Steuben and Rust Formations) is now shown to transition eastward across a series of faults that were activated syndepositionally. The fossiliferous subtidal limestone facies of the type Trenton Falls region changes rapidly into a very distinctive rhythmic-bedded, dark gray shale and calcisiltite unit, also referred to as “ribbon limestones.” This rhythmic unit, interpreted to represent turbidites deposited in deeper (and deepening) water on a steeply dipping ramp, is referred to as the Dolgeville Formation. As a unit, it is defined by the presence of *O. ruedemmani* zone graptolites and basal *C. spiniferus* zone graptolites (Goldman, 1993; Goldman et al, 1994). It appears that lower Trenton units have the widest distribution and lateral continuity of all the Trenton and extend as recognizable Trenton through the eastern half of the state where the equivalents are referred to as the Glens Falls Formation. In this region, the top contact of the Trenton is demarcated by the rapid upward shift to dark gray and black shales of the Flat Creek Formation (Brett & Baird, 2002). In the eastern Mohawk Valley region, black shale deposition commenced during the early *C. americanus* graptolite zone. Thus, the top contact of the Trenton in this region is likely the equivalent of the uppermost or Rathbun member of the Sugar River Formation.

To the northeast of the Trenton Falls region, in the northern Black River Valley, the uppermost unit of the Trenton – the Steuben equivalent – is overlain by another thin transitional limestone that resembles beds of the middle Trenton at Trenton Falls. This unit, named the Hillier (Kay, 1937), is recognized by the much darker colored limestones with shaly interbeds that are in stark contrast to the light colored fossiliferous grainstones of the Steuben (Hallowell of Kay, 1937). As elsewhere, and as typical for the upper Trenton contact in eastern New York, the contact at the top of the Trenton in this region is also marked by a sharp facies change into condensed, extremely organic-rich black shales interbedded with thin allodapic limestones that suggest a deepening motif followed by shallowing in the overlying shales. In central New York, as at Trenton Falls, these shales are referred to as the Indian Castle Shales. In the southern Black River Valley they have been referred to as the Holland Patent Shales, in the northern Black River Valley they have been referred to as the Deer River Shales and in Ontario – the same interval is referred to as the Collingwood Shale. These shales are so enriched in organic material they have been considered oil shales and are potentially viable for oil production (Dyni, 2006). Across the region, the shales contain *C. spiniferus* zone graptolites (in the L. Indian Castle Shales) followed by *G. pygmaeus* zone graptolites in the remainder of the units.

In the type region in the vicinity of Trenton Falls, the Trenton Limestone is shown to be almost 100 m (300') thick (Coryell, 1915). The Trenton Limestones reach their maximum thickness in a NE/SW trending depocenter focused in the vicinity of the Olin Basin which may extend further to the NE/SW to be coincident with the Kingston Trough (**see figure 12**). To the northwest along the margin of the Adirondacks and Frontenac Arch, the thickness of the Trenton rocks is somewhat difficult to establish due to subsequent erosion and stratal tilting; however, data from cores in the Lake Ontario region and outcrops along the northern margin of Lake Ontario show that the Trenton equivalents are thick in the vicinity of the Kingston Trough and then thin slightly in the Marmora region before they thicken again in the Lake Simcoe region. Liberty (1969) calculated the total thickness for the Trenton equivalents to be about 140 meters (420') in the latter region **(see figure 3)**. Farther west, into the Michigan Basin the Trenton equivalents thin once again and continue to thin through the Upper Mississippi Valley.

To the south of Ontario, and to the southwest of the type section, Trenton Limestone is shown to substantially thicken into southern New York and Pennsylvania where thicknesses exceed 200 m (600') along the Alleghany Front in the position of the Olin Basin (Fettke, 1948). To the east and south of the Olin Basin, Kay (1944), Thompson (1963), and Ryder (1991) document the thinning of the Trenton Limestones into the Appalachian Fold and Thrust across the Adirondack Arch. In this region, it is still not clear as to the specific thickness of the Trenton equivalent rocks because the rocks become dominated by siliciclastic beds as mentioned previously as it transitions into what has been referred to as the Martinsburg Group in that region.

In the type region, the Trenton Group is defined to contain eight internal subdivisions recognized on the basis of a range of bedding and sedimentation patterns (**figure 13**). Each is

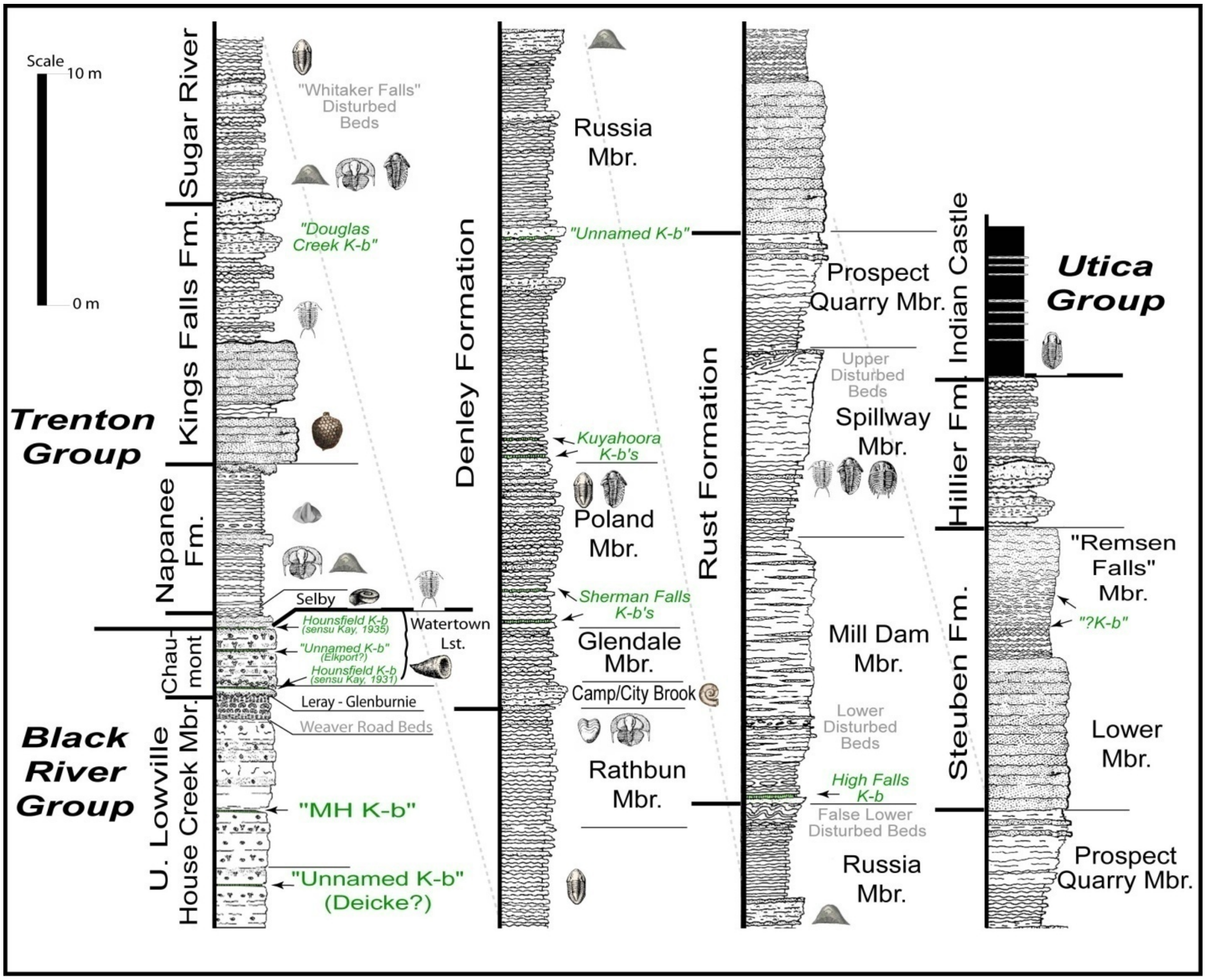


Figure 13: Composite stratigraphic column for the Trenton Groups in the Black River to Western Mohawk Valleys (modified after Cornell, 2005).

considered in turn in the following discussion.

Trenton Group Formations: Selby

In the Watertown, New York to Kingston, Ontario area, the massive and often coarse grained Watertown Limestone is overlain by about 0.5 to 3 m of nodular bedded, bituminous dark gray packstone to wackestone, termed the Selby. Although a complete outcrop section of the Selby is lacking in Ontario, this interval becomes substantially reduced and thins to a single condensed packstone to wackestone observed in localities south of Martinsburg, New York. The Selby Formation, originally defined by Marshall Kay (1937) for exposures along Selby Creek, Ontario, has historically represented the basal part of the Trenton Group and was deposited during early Rocklandian (**figure 14**). Kay grouped the Selby with the overlying Napanee

Figure 14: Satellite and topographic map overlay for the Napanee, Ontario Region. Shown on the map are the locations of the Napanee and Selby type sections. Map produced using images obtained from NASA World Wind digital globe.

Formation into his Rockland Formation. Since this assessment was established, the term Rockland has been designated as a time-rock term and the Selby Limestone was in turn promoted to formation status. The Selby Limestone includes the set of limestones between the Watertown Limestone and the Napanee Formation.

Lithologically, the Selby consists of light weathering limestones that on fresh surfaces are dark-gray to black in color. These beds are typically medium to fine-grained and often weather to a buff yellow in areas where the Selby contains slightly more magnesium-rich carbonates. Bedding is typically massive to bioturbated and nodular. In well weathered outcrops the Selby contrasts with the overlying Napanee in that it lacks the shaly interbeds typical of the Napanee.

Moreover, the Selby in most outcrops exhibits a condensed appearance with a low diversity, but high abundance of cephalopods including large endoceratids. The highly abundant paragastrapod *Maclurites logani* is also found in the dense accumulations of current-oriented orthoconic cephalopods. Collectively, they resemble a "cephalopod limestone" facies.

The Selby is usually a very thin unit and is not mapped or often exposed, however it does contain a rather homogeneous lithology where it is exposed and recognized. The upper contact of the Selby, beneath the Napanee Formation, is a characteristic rusty weathering iron-mineralized surface. In the Mohawk Valley region, where the Selby is mostly absent, one locality (at Ingham's Mills, on East Canada Creek) has a thin 40cm interval of Selby. In this location, the unit is riddled with distinctive vertical tubes between 0.5 and 1 cm in diameter that have been in filled with quartz, calcite, pyrite, feldspar, and organic carbon in the form of anthraxolite (Argast, 1992). These carbon-rich tube infillings have negative δ^{13}C ratios (-23.4 to -25.0). These carbon-filled tubes have also been identified in a number of cores further east at the same stratigraphic level (Baird & Brett, 2008). Although the exact origin of the infillings is currently unknown, they do appear to be related to condensation and sediment starvation at discontinuity surfaces.

Elsewhere an additional marker bed lies at the base of the formation. Near the top of the northern side of the power diversion channel for the Black River hydroelectric generation facility at Glenn Park, New York (near Watertown, NY) a prominent yellow-tan weathering clay seam is well exposed below the typical nodular thin bedded Selby. In this location the Selby is very rich in brachiopods including the strophomenid *Sowerbyella curdsvillensis.* The yellow colored K-bentonite, was named the Hounsfield K-bentonite (sensu Kay, 1935) after he had given the same name to a bentonite at a lower position (sensu Kay, 1931; **see figure 13**). Together with at least three other beds in the Watertown and Lowville formations below, this K-bentonite can be traced northwest into Ontario where they are well exposed at Napanee, Marmora, and in the Lake Simcoe region. Some of these K-bentonites are also traceable toward the southeast and although their presence is often difficult to detect without the aid of hand-held scintilometer, they have been recognized in the West Canada Creek Valley. In this region they are exposed in several stream and quarry sections in the vicinity of Middleville. As mentioned, the basal Selby K-bentonite was the focus of several papers by Kay (1931, 1935), and although there is some more recent controversy surrounding the name and correlation of this and subjacent K-bentonites (see Brett et al., 2004; Mitchell et al., 2004; Carey, 2006), the basal Selby K-bentonite at Middleville was referred to as the equivalent of the Hounsfield Metabentonite of the northern Black River Valley (sensu Kay, 1935).

Trenton Group Formations: Napanee Formation

As in most sections, especially to the northwest and southeast of Middleville, the Selby interval is overlain by thin bedded, platy calcisiltites, and interbedded wackestones and shales assigned to the Napanee Formation. Collectively, the Selby and Napanee by definition have been assigned to the Rocklandian Stage, based on lithologic and biostratigraphic evidence. Despite its rather distinctive lithologic appearance in many regions, the Napanee was first defined on the basis of its faunal composition. Originally, Kay (1937) designated this unit (the upper member of the Rockland) on the presence of the distinctive bi-lobed brachiopod *Triplecia cuspidata* (Hall). These *Triplecia*- bearing strata were superjacent to the Selby Limestone and subjacent to the Hull (or Kings Falls Limestones). In addition, the Napanee tends to be heavily dominated by the small orthid brachiopod *Paucicrura* (*Dalmanella*) *rogata* (Sardeson) and like the underlying Selby and Watertown contains the strophomenid brachiopod *Sowerbyella curdsvillensis*.

Lithologically, the Napanee Formation can be recognized independently of faunal evidence based on its stratigraphic position above the underlying Selby and beneath the overlying crinoidal grainstones of the Kings Falls Formation. Unfortunately owing to the softer weathering shaly interbeds, the Napanee is poorly exposed in most regions and is often weathered back along terrace plains and covered by colluvial debris. Some key sections can be observed including those at Sugar River near Boonville. Nonetheless, this unit tends to show rather rhythmic interbeds of fine grained calcilutites, calcisiltites and interbedded shales. Depending on location, the caps of calcilutite beds and the shales can be quite fossiliferous with brachiopods, crinoids, trilobites, various mollusks and trilobites including *Cryptolithus tesselatus* Green as at Ingham Mills. In weathered sections this unit tends to weather buff tan but in fresh surfaces, the medium to fine textured carbonates tend to be dark brownish gray in color.

The Napanee Formation was originally proposed by Kay (1937) to include the rock overlying the Black River Group in southeastern Ontario, near the town of Napanee, Ontario (**see figure 13**). This unit formed the upper portion of the Rockland Limestone (sensu Raymond, 1914; Rockland, Ontario). The Napanee varies in its thickness (from less than one meter to about eight meters), especially in the southern Black River and the western Mohawk valleys. Where the Napanee is truncated by the overlying Kings Falls Formation, it can be reduced to less than a few feet as at the south end of Middleville, New York. To the north, in its type region, it consists of its maximum thickness of about eight meters.

In most areas of northwestern New York and Ontario, Canada, the Napanee Formation tends to be rather homogenous in its lithologic character. However, despite its seemingly regular appearance, the Napanee does have a 0.5 to 1 meter thick middle unit that tends to be slightly more fossiliferous and can even develop into packstone to grainstone facies. In some instances brachiopods can show vertical orientation and/or imbrications suggesting. This middle unit can be used to divide the Napanee into a lower and upper unit (Cameron, 1968). In regard to other stratigraphic marker beds, the sharp basal contact of the Napanee with the underlying Selby tends to be its most striking feature. The dramatic shift from a dark gray massive burrow mottled wackestone and micritic packstones to alternating calcilutites, calcisiltites, and shales occurs at a very sharp pyrite and phosphate enriched surface. This surface can be traced some distance from Ontario into New York State. In the West Canada Creek to Mohawk Valleys the underlying Selby pinches out below this contact and as such, this contact has been variously interpreted as a subaerial erosion surface (Cameron & Mangion, 1977) or as a maximum flooding surface (Cornell, 2001). In addition to lithologic characteristics, there are at least two persistent zones that show an abundance of *Triplecia* and may represent acme zones or epiboles. In any case, these have been useful in mapping in the central Black River Valley from Deer River through at least Boonville. Their correlation into the western Mohawk Valley region is not confirmed, however, there is at least one zone of these brachiopods found at Ingham's Mills.

Trenton Group Formations: Kings Falls Formation

Kay (1968) introduced the term Kings Falls Formation, for the rocks equivalent to the Hull and Kirkfield Limestones found in Ontario. This unit is coarse-grained and dominated by medium-bedded brachiopod-crinoid grainstones and skeletal packstones interbedded with thin stringers of shale. Kay had chosen to use Kirkfield as a time-rock term (i.e. Kirkfieldian Stage). In order to establish a rock-term to use for these strata; he selected the best exposure of the basal

Trenton in the northern New York region. For the type section of these limestones, Kay chose Kings Falls downstream from Copenhagen, NY on the Deer River (**figure 15**). This tributary

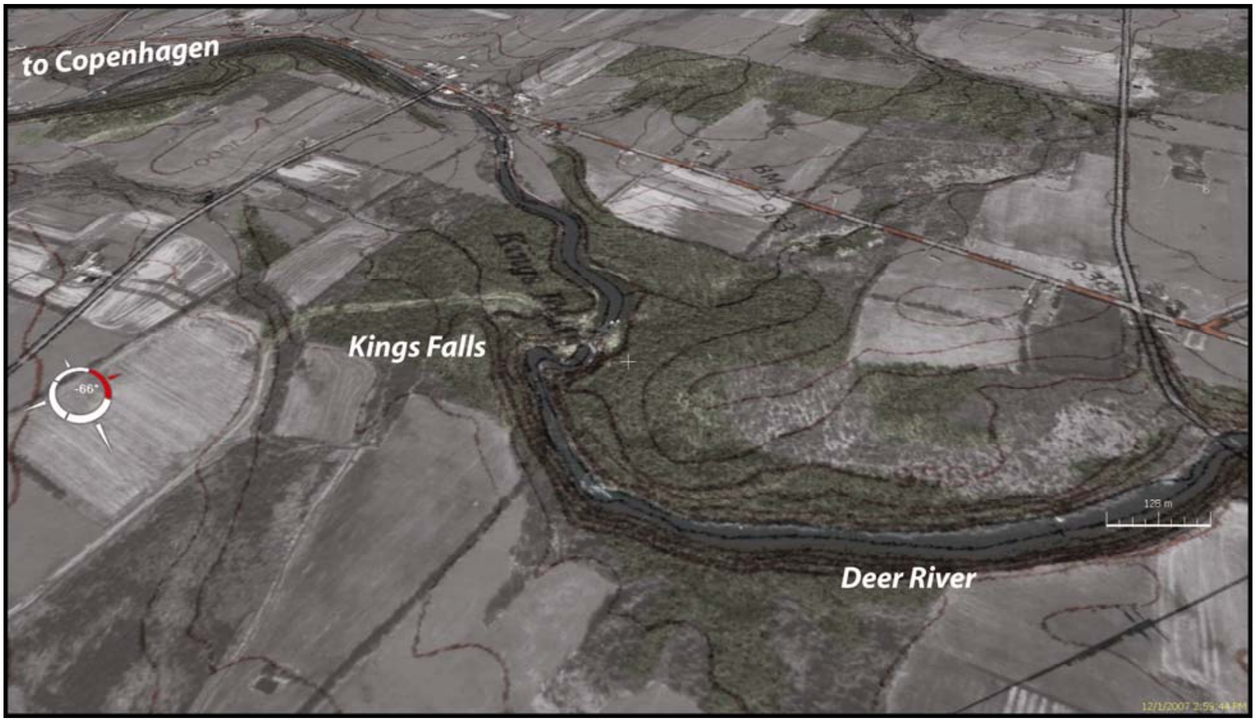


Figure 15: Type locality for the Kings Falls Formation on Deer River, east of Copenhagen, New York. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the northwest.

drains the northern edge of the Tug Hill Plateau and joins the Black River just northwest of Lowville. The Kings Falls locality exposes both the basal contact with the underlying Napanee limestone, as well as the overlying Sugar River beds.

The contact between the Napanee and the overlying Kings Falls Formation is very sharp and often dramatic in weathered outcrop sections. In these cases the Kings Falls often ledges out over the Napanee. In some places channels and erosional features have been shown to truncate upper Napanee beds. This is evident at several localities including those at Ingham Mills, and Middleville in the Mohawk Valley and at Boonville on the Sugar River.

The upper contact of the Kings Falls was marked by the highest occurrence of well-bedded grainstone-packstone coquina beds (Kay, 1968). Although a lithologic transition into finer grained nodular wackestones and fine grained grainstones actually begins below this position, the contact was drawn near the top of the falls at the type section on Deer River. The contact is recognized below the cap of Kings Falls on Deer River, and below the base of a distinctive deformed interval. The deformed beds are exposed both at Deer River (where there is evidence to support synsedimentary movement on local faults) and in the wall below the lower Whitaker Falls on Roaring Brook.

Nonetheless, the highest occurrence of coarse grainstones of Kings Falls lithology (as defined by Kay) oversteps a hardground surface now recognized as the base of the Sugar River Formation at the type section on Sugar River at Boonville. Herein, it is proposed that the Kings Falls – Sugar River contact of Kay (1968) be lowered slightly to the contact recognized by Chenoweth (1952) as the base of the Shoreham Formation. This contact is not only recognized as a fairly well-developed undulatory hardground, but it also demarcates a characteristic change in bedding style from thicker beds of the Kings Falls to much thinner-bedded nodular fine-grained grainstones interbedded with calcareous dark shales. In many outcrops, especially in the Watertown region, the beds above the Kings Falls weather buff to dark brown and may show significant iron oxide development on some bedding planes.

As a unit then, the Kings Falls Formation is dominated by medium to thick bedded, cross-stratified, pelmatozoan grainstones especially in the base. These basal beds grade upward into mega-rippled grainstones. These grainstones transition to packstones and show increasing thicknesses of shale and slightly deeper water taxa. Within this overall deepening trend, the lower portion of the Kings Falls contains several meter-scale shallowing-upward cycles of thick ledges capped by calcarenites or crinoidal grainstones and brachiopod coquinas. Most often the grainstones show excellently preserved hummocky cross-stratification, and large-scale pararipples on the top surfaces of many beds. As mentioned the distinctive component of this particular interval is the dominance of echinoderm skeletal remains. In total the Kings Falls Formation ranges in thickness from near extinction on the Canajoharie Arch to nearly 35 meters near Sackets Harbor on the south shore of Lake Ontario west of Watertown, New York.

Interbedded with these very coarse-grained lithologies are a series of thin medium to dark gray shales. The shales often weather rusty, showing some evidence for pyrite development, and often contain excellently preserved trilobite remains including *Ceraurus pleurexanthemus* Green. In the Ontario region, this unit is equivalent to the Kirkfield Member (of the uppermost Bobcaygeon Formation) and is demonstrated to contain many classes of primitive and advanced echinoderms including asteroids, ophiuroids, rhombiferan cystoids, edrioasteroids, carpoids, crinoids, paracrinoids, and potentially even early echinoids (Brett and Taylor, 1999; Brett and Rudkin, 1997). While most of these echinoderms have not been recorded from entire specimens in New York, numerous plates are found that appear to be derived from many of these taxa.

In terms of key marker horizons, when compared to underlying and overlying units, the Kings Falls interval has relatively few K-bentonite markers for correlation. Nonetheless, Chenoweth (1952) indicates that there are occasional metabentonites in some sections but he did not reliably correlate them. He recognized a prominent clay seam near the top of the Kings Falls that was commonly about 1” thick. At Rathbun Brook northwest of Newport, NY in the West Canada Creek Valley, this seam is approximately one meter below the top contact. Chenoweth also located this same bed in at least three locations between Sugar River and Kings Falls in the Black River Valley. In this region it is consistently located just over one meter below the top contact. Herein it is referred to informally as the Douglas Creek K-bentonite (**see figure 13**).

Farther north, Chenoweth (1952) found a K-bentonite seam that was located almost five meters below the top of the Kings Falls in the vicinity of Carthage and another at nearly seven meters below the top of the Kings Falls at Watertown. He does not correlate this bed in the northern Black River Valley with that in the southern Black River Valley. He does, however, indicate that a seam is found nearly four meters below the top contact at a locality on Stony Creek south of Kings Falls. Although not stated, either a single seam is 1) correlative between all sections showing some variability in its vertical position with respect to the top contact owing to paleotopographic highs/lows; or 2) there are two distinct K-bentonites one near the summit of the Kings Falls Formation – the Douglas Creek, and another some distance below. At the current time, these hypotheses have not been tested.

Nonetheless, the upper surface of the Kings Falls tends to have a very sharp mineralized contact that is developed into an excellent hardground surface and although cryptic in some sections, it is distinctive and recognizable in many locations. In northern New York this upper surface of the Kings Falls tends to show a wide variety of sedimentologic and faunal components that make it a very unique surface. In most localities in the Lake Simcoe region of Ontario, as well as in the region south of Marmora, the basal beds of the equivalent Verulam Formation contain a series of hardgrounds and or firm grounds and tend to be densely colonized by several species of bryozoans and are commonly bored by *Trypanites* (Brett and Liddell, 1978). The uppermost Kings Falls hardground is not recognized south of the Black River Valley primarily due to the lack of exposure in this interval. However in sections in the Newport to Middleville region along West Canada Creek, as at Rathbun Brook and again at City Brook, there are several closely spaced hardground surfaces that are likely candidates. The uppermost of these is overlain with a well developed dark shale interval at the base of which is a very dense accumulation of especially large *Prasopora simulatrix* bryozoan colonies. These colonies are often several inches in diameter and many of them show evidence of reworking, abrasion, and some show *Trypanites* borings. Although it is too premature to establish this correlation confidently, the presence of the subjacent clay seam recognized by Chenoweth (1952) helps provide additional support for this potential assessment.

In addition to its upper hardground bed, the Kings Falls Limestone contains several additional hardgrounds within its lower half. At Boonville, and again in Watertown, N.Y., several prominent hardground surfaces are recognized and contain rip-up clasts, highly abraded fossil fragments, and in some cases there are reworked pebbles and cephalopod steinkerns. To date, few articulated echinoderm taxa are known from these hardgrounds. However, hardgrounds in this same interval within the upper Bobcaygeon of the Lake Simcoe region of Ontario are well-known for their echinoderm faunas (Brett & Brookfield, 1984; Brett & Taylor, 1999; Sumrall & Gahn, 2006). They are also known to contain several intraformational conglomerates (Liberty, 1969; Cornell, 2001). Although individual hardground beds have not yet been confidently correlated across the entire region, small-scale cycles within major facies packages in this interval are fairly easily recognized and have provided a basis for strengthening the correlations of these units.

In the Mohawk River Valley, the lower Kings Falls Formation at Middleville and again at Ingham's Mills on East Canada Creek contains a series of two to three basal polymictic conglomerate beds. At Middleville, the conglomerate beds are mostly made of limestone intraclasts as is the case to the northwest. However the limestones do have occasional quartz pebbles making them at least diamictic. At Ingham's, in addition to limestone intraclasts, these beds contain granitic gneiss clasts admixed with several lithologies of limestones and quartz pebbles. Combined with evidence of truncation of subjacent Napanee Limestones in several locations in this region, the presence of these intraclasts (including siliciclastic and metamorphic) suggest non-uniform topographic inversion associated with fault movements in this region at this time.

<u>Trenton Group Formations: Sugar River Formation</u>

In the Black River Valley region and in the nearby West Canada Creek Valley, the upper Kings Falls changes rapidly into thin interbedded shales, fine-grainstones, and nodular wackestones. In some areas calcisiltite stringers become common. This lithology is more similar to the Napanee Formation subjacent to the Kings Falls, than it is to the coarser-grained Kings Falls, but because of its nodular bedding it can be challenging to identify the specific base of the unit – especially where the Kings Falls is not well-exposed. The Sugar River Limestone was first established as a rock-term to replace the term Shoreham Limestone, which Kay (1968) had elevated to sub-stage status (i.e. Shorehamian substage of the Shermanian Stage). Similar to the Kings Falls, the Sugar River was described from a locality in Lewis County, New York in the upper Black River Valley near its drainage divide with West Canada Creek (**figure 16**).

Although the Sugar River is dominantly interbedded shales, packstones, and grainstones, it becomes distinctively more nodular and finer-grained towards the middle of the unit. In the type section at Sugar River, southern Lewis County, the formation is approximately 12 meters (40') thick. The middle Sugar River grades into a more medium-bedded, calcisiltite facies. Beds show graded bedding and generally contain fewer fossiliferous horizons although *Rafinesquina deltoidea* brachiopod coquinas are common on some bedding planes near the top of the unit. This upper unit has been recognized as a separate unit and named the Rathbun Member (Kay, 1943; Chenoweth, 1952).

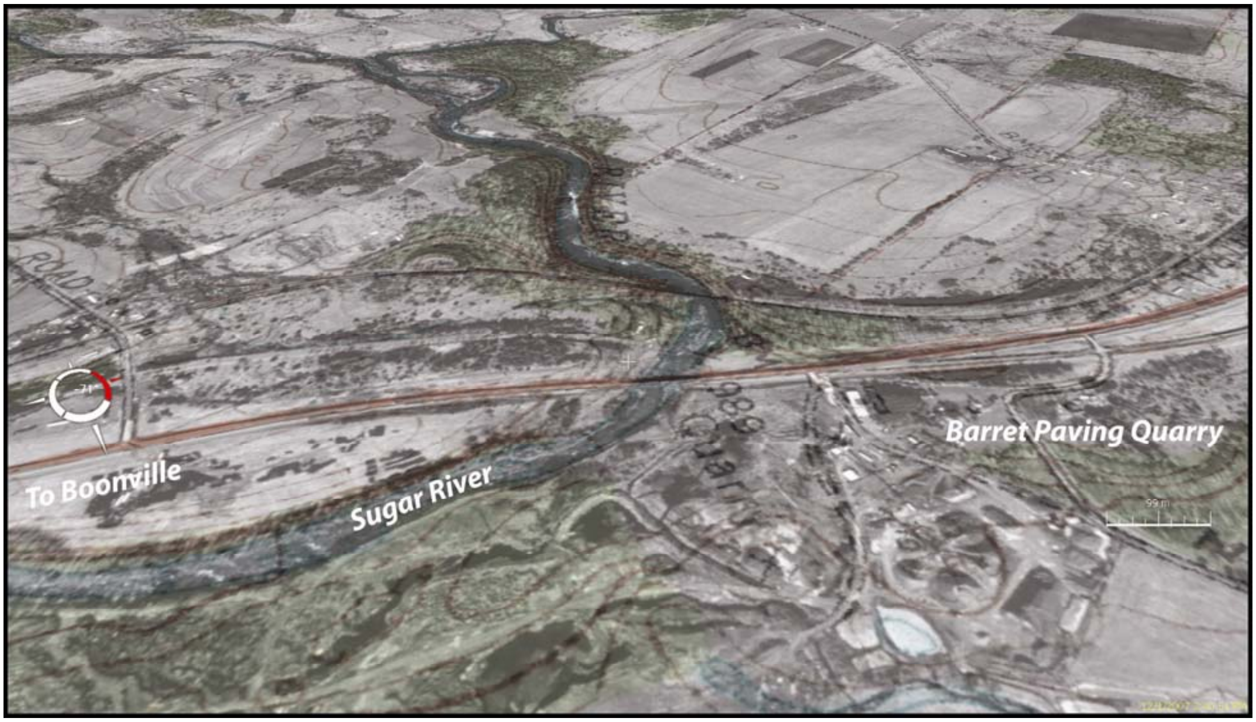


Figure 16: Type locality for the Sugar River Formation on the Sugar River, north of Boonville, New York. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the west by northwest.

To the southeast of the type Sugar River sections, the Rathbun Member is traceable into the West Canada Creek region where it was first defined by Kay. In the Rathbun Brook type section, the upper Sugar River becomes even finer-grained, less nodular, and has well-developed calcilutites. In a very short distance to the south near Middleville, New York the lutites have been shown to contain crinoid and brachiopod stringers with occasional grainstone stringers. Thus south of Sugar River, this upper member shows evidence of lateral deepening followed by rapid shallowing and thinning in the West Canada Creek Valley before it transitions into the shale dominated facies of the lower Flat Creek Shales in the eastern Mohawk Valley east of Little Falls. To the northwest of Sugar River, Chenoweth (1952) was able to identify the distinctive facies of the upper Sugar River Rathbun member as far north as Mill Creek on the south shore of Lake Ontario near Sackets Harbor. In the West Canada Creek Valley, the Sugar River has been measured at approximately ten meters while in northern New York it is nearly the same thickness. However, in the southern Black River Valley region, the Sugar River Formation reaches a maximum thickness near the type section at Sugar River where it is nearly seventeen meters (Chenoweth, 1952).

In addition to its distinctive lithology, the Sugar River carries a very distinctive fauna that enables its easy field recognition wherever it is exposed. This interval contains a well-known acme zone of two major faunal elements in the New York sections. First, the lace-collared trilobite *Cryptolithus tesselatus* (Green) is present in large numbers at several horizons, but most predominantly in the medial Sugar River. Second, large gumdrop-shaped bryozoans, *Prasopora* occurs abundantly within the lower portion of the Sugar River. Both of these faunal elements are abundant and often compose very dense accumulations on some bedding planes as mentioned. Otherwise, the Sugar River Formation demonstrates few important internal marker beds itself that are useful in correlation. The exceptions are the key contacts at its base, the distinctive Rathbun interval, and its upper contact with the overlying Denley Formation. There is one further internal interval that has the potential for further consideration and investigation. In the lower portion of the Sugar River Formation at Kings Falls just below the cap of the falls is an interval of contorted strata referred to by Chenoweth (1952) as “subsolifluction structures” which are subaqueous sediments sliding on low gradients. These contorted strata show evidence of channeling, disc and saucer structures as well as shale diapirs that are very similar to structures found in the Lexington Limestone of Kentucky.

The latter have been interpreted as seismites (Ettensohn et al., 2002). The presence of these features indicates possible synsedimentary deformation of sediments on the seafloor perhaps due to local fault activity. Similar structures are observed in the same stratigraphic position at Whitaker Falls in Martinsburg, New York and elsewhere in the central to northern

Black River Valley (Chenoweth, 1952). Given the increased thickness of the Sugar River in this region of the Black River Valley it is plausible that local faults moved in response to Vermontian phase tectonism. Fault activity may have triggered uplift and subsidence on localized blocks especially to the east of the Trenton Falls region in the Mohawk Valley where Trenton equivalents become much more condensed and where black shale deposition begins in earnest in eastern New York State. In the Black River Valley this fault activity may have generated earthquake activity and/or initiated movement of sediments on the seafloor during this time.

Trenton Group Formations: Denley Formation

Although it has been recognized as a very distinctive stratigraphic interval since the first geologic investigations of Trenton Falls, the interval from near the cap of Sherman Fall upward to the base of the coarse-grained carbonate at the top of the gorge was not fully differentiated until the studies of Kay (1943, 1968). Initially, Kay applied the term Denmark Formation to this interval of dominantly fine-grained carbonates and interbedded shales, for exposures of the unit in the central Black River Valley in the town of Denmark. In 1968, Kay relegated the term Denmark for biostratigraphic purposes and applied the term Denley Formation as a rock terms. The type section of the Denley Formation is located upstream from the Sugar River type locality and in the tributary stream to the Sugar River called Moose Creek (**figure 17**).

In common usage, the Denley Formation is recognized lithologically on the occurrence of distinctive marker horizons at the lower and upper contacts of the formation. These two marker horizons include 1) the basal "*Trocholites* beds" of the Camp Member (Chenoweth, 1952) also referred to as the City Brook Beds (Baird et al., 1992; Brett & Baird, 2002) and 2) the High Falls K-bentonite from Trenton Falls (**figure 18**). Thus delineated, the Denley Formation now

Figure 17: Type locality for the Denley Formation on the Sugar River, north of Boonville, New York adjacent to Denley Road above Denley Station and along Moose Creek a tributary to the Sugar River. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the west.

encompasses only a portion of the original Denmark Formation. It is now defined by Brett and Baird (2002) to have two members: the lower or Poland Member and the upper or Russia Member, both of which are also names used by Kay.

The Denley is composed dominantly of fossiliferous nodular fine-grained limestones which show evidence for shallowing-upward cycle development and numerous internal and distinctive marker horizons. Each member of the Denley and their internal marker divisions are discussed in stratigraphic order (oldest first) in the following discussion.

Poland Member

The Poland Member is now defined between the basal City Brook or Camp sub-member and the lowermost Kuyahoora K-bentonite (Brett and Baird, 2002). As bracketed this member is a relatively condensed, nodular calcilutite interval containing *Trocholites* and transitions to more tabular-bedded barren calcilutites interbedded with buff weathering shales and thin coquinal beds

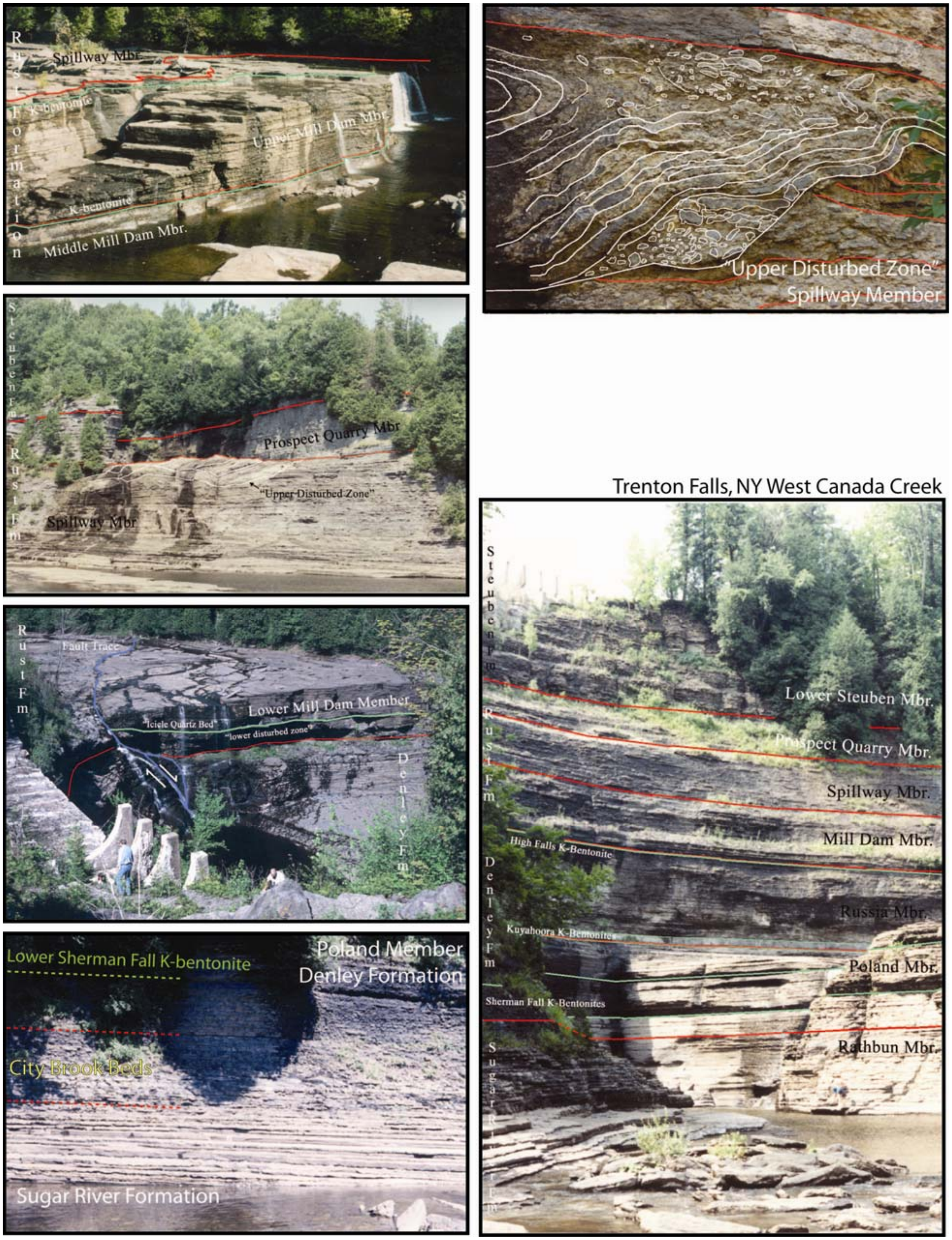


Figure 18: Outcrop images from the type Trenton Falls region on West Canada Creek, between Trenton Falls and Remsen Falls.

of the Glendale sub-member (upper Poland). This unit is moderately fossiliferous with species of trilobites especially *Isotelus*. The interval is relatively resistant to weathering, and makes up the ledge forming the cap of many waterfalls in the region including Sherman Fall at Trenton Falls.

Given this delineation, the Poland is approximately nine meters (30') thick. The City Brook beds, also referred to as the "Camp Member" by Chenoweth (1952), is a relatively condensed, nodular calcilutite interval containing abundant endocerids, nautiloids, and occasional *Trocholites sp.*. This interval measures two meters at Trenton Falls and thins to the north into the Black River Valley where it is less than one meter at Lowville. It again thickens substantially to almost four meters toward at the type locality of the Camp member near Lake Ontario (**figure 19**). The base of this unit is marked by a sharp discontinuity surface with the

Figure 19: Type locality for the Camp sub-member of the Poland Formation along Mill Creek, adjacent just east of Sackets Harbor, New York. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the east-southeast.

underlying Rathbun. Its upper limit is marked by the transition to more tabular-bedded barren calcilutites which are interbedded with buff weathering shales and thin coquinal grainstone beds of the Glendale sub-member. This latter interval, grades into planar-bedded calcisiltites and calcilutites with buff weathering shale seams. This particular interval, called the Glendale sub-member, is approximately eleven meters (35') thick in the southern Black River Valley and thins to less than eight meters (25') at Trenton Falls. This unit is fossiliferous with *Isotelus* trilobites especially common. The planar middle to upper Poland interval grades upward into coarser-grained, rippled calcarenites that often contain intraclasts showing evidence for early cementation and development as storm clasts, including in the type area (**figure 20**).

Figure 20: Type locality for the Glendale sub-member of the Poland Formation along Whetstone Creek, adjacent just east of Sackets Harbor, New York. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the east-southeast.

As previously mentioned the Poland Member is a distinctive interval that is exceedingly useful for correlation. Due to the distinctive Camp/City Brook Bed and the planar-bedded calcilutites, shales and calcarenites of the overlying Glendale sub-member, this interval is generally easily identified in outcrop. In addition, the top of the Glendale sub-member is relatively distinctive because of the presence of several intraclastic conglomerates within the wavy-bedded calcarenite interval in many locations. Adding to its distinctive lithologic character, this unit has excellently preserved *Isotelus* sp. and *Flexicalymene* sp. trilobites in some layers. In addition to these highly favored faunal elements, the Poland member contains typical Trenton faunas such as the brachiopods *Sowerbyella*, and *Rafinesquina* in very large numbers. Moreover, the presence of the graptolite *Corynoides americanus* biostratigraphically places this unit in the *C. americanus* zone.

Additional marker horizons include two K-bentonites in addition to the lower Kuyahoora K-bentonite located at the top of the Poland (Brett & Baird, 2002). These K-bentonite horizons are present in the face of Sherman Falls as recessive notches. The lower Sherman Fall K-bentonite is located about 4.5 meters above the base of the Glendale sub-member and has been chemically fingerprinted and correlated into equivalent units to the east of Trenton Falls. The upper Sherman Fall K-bentonite, although not yet fingerprinted, is located about three meters above the former. These twin K-bentonites can be used for correlation and were called the M8 and M11 K-bentonites by Cisne and Rabe (1978) and Cisne et al. (1982).

Russia Member:

The Russia member of the Denley Formation was originally named by Kay in (1943) for the upper calcilutite-dominated facies above the Poland and below the coarse-grained carbonates of the Cobourg (now called Rust Formation; Brett and Baird, 2002; **figure 21**). The unit extended upward to the base of the coarse-grained carbonates of the overlying Cobourg (now called Rust Formation). This particular member of the Denley is very distinct in that it possesses several

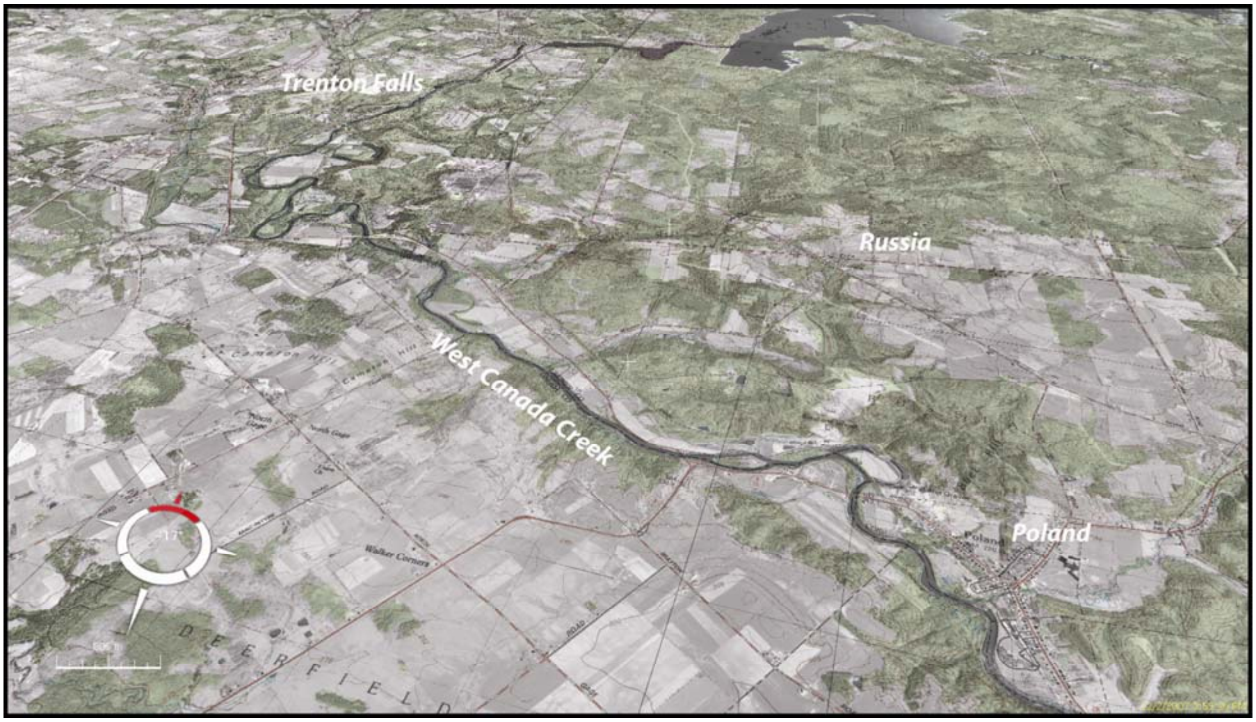


Figure 21: Type region for the Poland and Russia members of the Denley Formation in the West Canada Valley just southeast of Trenton Falls. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the north-northwest.

well-developed cycles, which show very distinctive shallowing upward motifs. As these cycles are very well developed and easily recognized, each has been named as an informal subunit of the Russia by Brett & Baird (2002). These cycles each in turn show slight facies differences between the lower, middle, and uppermost beds (showing a larger-scale shallowing upward pattern followed by increased condensation). As such three sub-members have been designated for the Russia similar to the submembers for the Poland, and five "cycles" have also been identified. The lower sub-member is called the Lower High Falls sub-member, the middle sub-member the Cincinnati Creek sub-member, and the upper the Upper High Falls sub-member.

The Lower High Falls sub-member includes the "Overhanging Ledge Bed" cycle and is the basal shallowing-upward cycle of the Russia. It is one of the most easily recognized of the five Russia cycles. It is denoted at its base by the contact with the subjacent Poland at the position of the Lower Kuyahoora K-bentonite (**figure 22**). The base of the unit also shows a

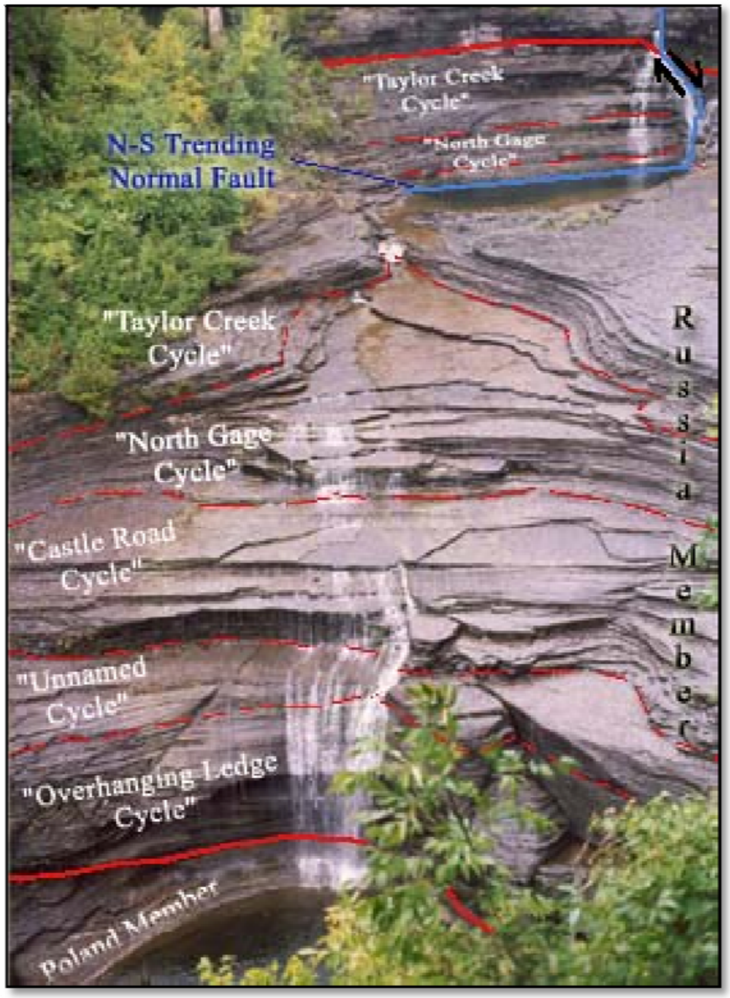


Figure 22: Exposures of the Russia Limestone member of the Denley Formation at Trenton Falls. Shown in the image are the Lower High Falls of Trenton gorge ranging up through the base of the Upper High Falls. Delineated in the image are the main cycles recognized by Brett & Baird (2002) and included in the 3 sub-members of the Russia.

sharp lithologic change from the rippled calcarenites and intraclastic limestones of the underlying Poland into rather homogeneous shaly-nodular fine-grained calcilutite limestones. Within the basal few feet of the lower cycle, a very weak sub-cycle can sometimes be identified between the Lower and Upper Kuyahoora K-bentonites with a somewhat condensed skeletal packstone at the top. Regardless, this cycle grades back into shaly nodular fine-grained limestone that persists upward to a compact coarse-grained crinoidal packstone bed that composes the lip of Lower High Falls. This sub-member's total thickness is then just about eight meters (25').

The middle or Cincinnati Creek sub-member is composed of two cycles a lower "unnamed" cycle and the "Castle Road Cycle." The "Castle Road Bed Cycle" is a substantially thinner cycle (~4.5 m), than underlying cycles. Yet because of its distinctive basal and capping beds, it is easily recognized. It has nearly 1' of dark gray to black fissile shale overlain by thin concretionary calcisiltites which grade upward through interbedded shales, and calcisiltites into a series of compact, condensed crinoidal wackestones and packstones. These mark the cap of the cycle and are referred to as the Castle Road Beds from exposures on nearby City Brook. Interrupting this succession is a minor persistent packstone to grainstone bed that is yet "unnamed," and is present in the middle part of this cycle. However, the Castle Road cycle cap is perhaps the most coarse-grained of the Russia and can be correlated some distance to the southeast of Trenton Falls. This cycle is also recognized due to the presence of an additional unnamed K-bentonite located immediately below the base of the capping bed. This K-bentonite has been found in several outcrops to the southeast of Trenton Falls.

Overlying the Castle Road Cycle is another slightly thicker cycle measuring approximately 5.5 meters thick. This cycle, called the "North Gage Cycle" is developed in dark gray shales interbedded with wavy-bedded calcisiltites which are capped by a very distinctive bed. This capping bed is widespread, and composed of a wavy-bedded *Prasopora* packstone or biorudite and marks the top of the Cincinnati Creek sub-member. This bed and a subjacent one meter (3') thick bioturbated, fine-grained grainstone bed make a distinctive couplet of beds that are easily recognized and traced widely in the West Canada Creek exposures.

The uppermost sub-member, the Upper High Falls interval, is recognized by a distinctive change in lithofacies out of the underlying Cincinnati Creek sub-member. The Taylor Fork Cycle which marks this change is approximately four meters thick with two meters at its base composed of tabular and even-bedded calcilutites. These beds grade upward into standard Russia-type nodular wackestones and packstones with well-developed bioturbation. The cap of this cycle is rather distinctive in that the uppermost beds show evidence of soft-sediment deformation. This interval has been referred to mistakenly as the "Lower Disturbed Zone" which occurs in the base of the overlying Rust Formation. Like those of the underlying Sugar

River, these beds suggest that soft-sediment deformation processes were again active in the region. Immediately overlying the Cincinnati Creek sub-member disturbed beds is a condensed, quartz-rich, crinoidal grainstone that truncates portions of the underlying disturbed interval. The top surface of this bed shows substantial mineralization including silicification, phosphatization, and dissolution of low Mg-calcite from skeletal fragments. Sitting on this contact is the Upper High Falls K-bentonite that demarcates the contact between the Russia Member of the Denley Formation and the superjacent Rust Formation.

Due in part to its prominent cyclicity and fossils, the Russia Member is a rather distinct stratigraphic interval within the middle Trenton Group. With unique event beds, faunal epiboles, and distinctive lithologic changes in key intervals, the Russia provides a range of excellent markers for recognition and correlation of the Russia Member into other areas. The series of marker horizons, although previously noted in the discussion of each of the sub-members, include metamorphosed volcanic ash layers: the Upper High Falls K-bentonite, the twin Kuyahoora K-bentonites, and an “unnamed” K-bentonite in the Cincinnati Creek sub-member. Likewise, the disturbed zone just below the cap of the Taylor Fork cycle represents a similar type of event horizon that helps establish a fairly distinctive time marker. Other marker horizons include the small-scale cycles themselves, as well as larger-scale sequence stratigraphic marker horizons that will be discussed in the next chapter. In addition, the Russia displays several faunal signatures that are constrained to characteristic stratigraphic horizons. As discussed, the couplet beds of the upper North Gage Road cycle are unique in their distinctive bioturbated, fine-grained grainstone and *Prasopora* biorudites. These two beds are utterly unique and easily recognized over wide-regions within this interval. The development of this interval is likely due to unique environmental conditions that existed for a very short time during the deposition of the upper Russia.

Trenton Group Formations: Rust Formation

Kay (1943) introduced the term Rust for the lower member of the Cobourg Formation. He defined two divisions in West Canada Creek. He named the lower unit the Rust, after exposures in the quarries on the William Rust Farm just east of Trenton Falls (**figure 23**). Up to that point, the interval near the top of the Trenton was not differentiated, but because of its rather

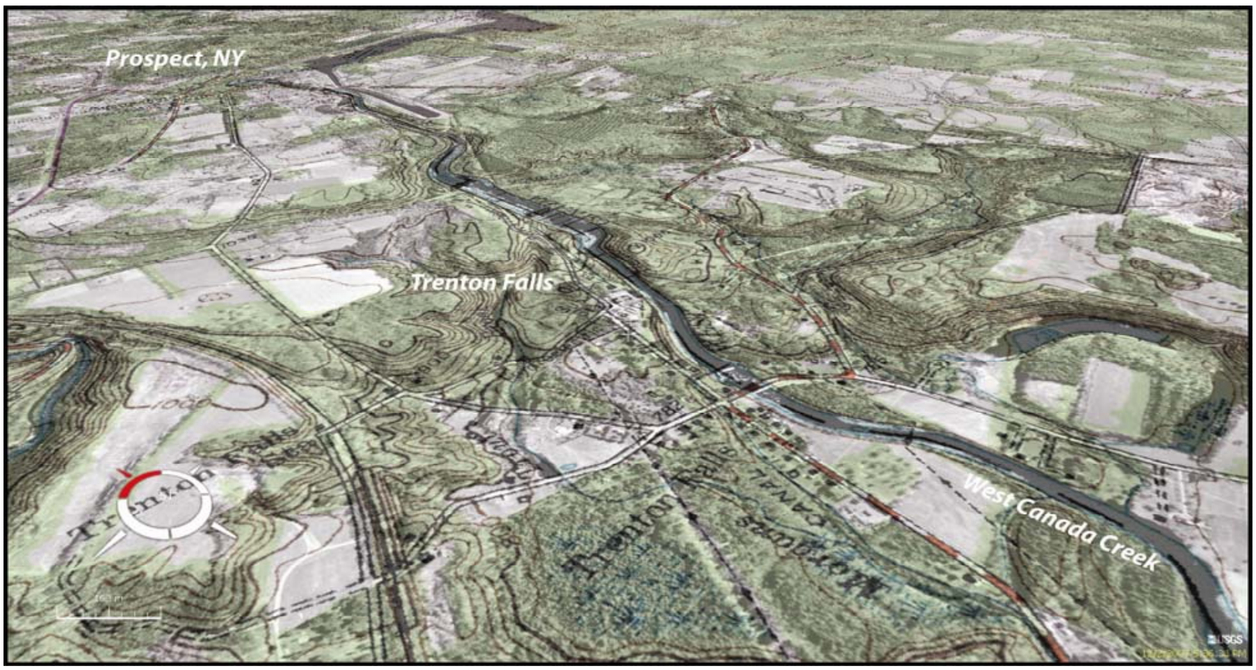


Figure 23: Type region for the Trenton Group in the West Canada Valley. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the northeast.

distinctive facies, it was separated from the overlying Steuben or upper Cobourg. Recent review by Brett & Baird, (2002) has redefined the nature of the Rust such that, in contrast to the finer-grained Russia below, this unit is generally considered to be composed of nodular to wavy-bedded, coarse-grained packstones to grainstones. These beds can internally be divided into three lithologic divisions. Although very similar in overall composition, these divisions can be distinguished as shallowing-upward packages with similar motifs to the underlying Russia. Due to its coarser lithology, the Rust is very easily recognized in the West Canada Creek Valley. Farther to the east, however, correlations show that this interval becomes substantially modified and grades into a condensed turbidite-shale succession and then finally into basinal black Flat Creek shales (Valley Brook member). Internally, there are many interesting horizons useful in correlation including K-bentonites, and deformed or "disturbed" horizons.

In addition to lithologic characteristics, the Rust Formation is one of the best-known stratigraphic intervals in the Trenton Group because of the well-known collection of trilobites, echinoderms, and other fossil specimens collected from this interval, by William Rust and Charles Doolittle Walcott. With respect to biostratigraphic zonation, the boundary between the *Corynoides americanus* and *Orthograptus ruedemanni* graptolite zones is now recognized to occur within the base of the Rust Formation (Mitchell et al., 2002; Joy et al., 2000).

As mentioned above, the Rust Formation can be divided into three members: the Mill Dam member, the Spillway member, and the uppermost Prospect Quarry member. Each member and their internal marker divisions are discussed in stratigraphic order in the following section.

Mill Dam Member

The basal member of the Rust is represented by the twelve meters of rock above the Upper High Falls K-bentonite. The basal 0.5 meter of the Mill Dam, is similar in lithology to the Russia and is composed of tabular lutites and shales overlain by much coarser, and more resistant fossiliferous packstone beds. These rocks are exposed in the face of Upper High Falls and again in the interval immediately above the High Falls through the top of Mill Dam Fall due to faulting.

Just above the basal lutite package and within the base of the massive packstone beds is an important marker horizon, referred to informally by previous workers including Kay (1953), as the “Lower Disturbed Zone.” This distinctive, strongly deformed and contorted interval studied since the turn of the century (see Hahn 1913; Miller, 1915), occurs two meters above the base of the Mill Dam Member. The remainder of the Mill Dam Member is composed of a rather massive succession cyclic, wavy-bedded packstones interrupted by a thin interval of more tabular fine-grained wackestones. At Trenton Falls, the tabular fine-grained wackestone interval contains a thin recessive weathering notch that potentially represents another K-bentonite (Brett & Baird, 2002). The bench of limestone above Mill Dam Fall demarcates the cap of the Mill Dam Member, and is represented by a fairly compact interval of grainstones. The cap of these grainstone beds is very sharp with a change into shalier fine-grained lithologies of the overlying Rust Quarry Beds.

Spillway Member

The middle or Spillway Member of the Rust Formation is exposed from the base of the Hydro-dam Spillway ledge upward to the floor of the spillway. The spillway is located on the east side of the gorge just upstream from the railroad bridge crossing West Canada Creek. In its entirety, the Spillway Member is roughly the same thickness as the underlying Mill Dam Member (~twelve meters), but shows a greater degree of facies heterogeneity from base to summit. The lower portion of limestone at the base of the spillway ledge is identified as the same facies that occur in the Rust Quarry. In the Rust quarry, these beds are usually sparsely fossiliferous calcilutites; however, some beds contain the spectacular Walcott-Rust fossil occurrences. Thus they have been referred to as the Rust Quarry sub-member (Brett & Baird, 2002). The upper portion of the spillway interval returns to a more distinctive brachiopod-rich, wavy-bedded packstone interval that extends from midway up the face to the floor of the waterfall.

Of all Trenton intervals, the paleontology of the Rust Quarry Beds are perhaps the most well-known (**see figure 18**). Exposures of this interval occur upstream of Trenton Falls, near Prospect under the Military Road Bridge and in the Rust-Walcott Quarry just east of the gorge. The fine-grained calcilutite and shales have yielded some of the most-diverse, well-preserved fossil trilobites and echinoderms known from the Trenton (Delo, 1934; Brett et al. 1999). Sedimentologically, this unit fines upward through a repetitious series of fine-grained normal to reverse-graded peloidal calcilutites, calcisiltites and fine-grained calcarenites. The unique taphonomic signature of the lower Rust Quarry interval makes a useful marker in the base of the Spillway Member. This interval, just under a one meter-thick, displays uniquely preserved trilobites of at least eighteen species (some of which have preserved appendages). The preservation of these specimens and their taphonomic condition indicate that these trilobites were transported and buried alive by distal storm and turbidite flows. The interval also lacks significant evidence for bioturbation which helped to preserve these specimens (Brett et al. 1999).

The upper beds of the Spillway Member are very distinctive in their sedimentary character. They show an upward coarsening package from the lutites of the Rust Quarry interval through approximately five meters. The upward coarsening continues to the floor of the Spillway where the upper meter of coarse brachiopod packstone to grainstone beds is heavily deformed and slumped into broad channel-shaped features that have a nearly east to west, long-axis orientation. This interval has been referred to as the “Upper Disturbed Zone” (Kay, 1953) and has been recognized by previous authors (Sherman, 1826; White, 1896; Hahn, 1913; Miller, 1915), but has not been studied recently.

As mentioned, the Spillway Member succession is excellently constrained above and below by sharp flooding style contacts: Mill Dam Member below and by the shaly, interbedded calcisiltites of the overlying Prospect Quarry Member. These two contacts are recognized by the change from coarse-grained pack- to grainstone beds into shaly calcilutite to calcisiltite facies. The upper contact of the Spillway Member and the occurrence of the "Upper Disturbed Zone," makes this contact even more conspicuous at many localities in the West Canada Creek Valley. A condensed phosphatic lag is developed immediately on top of the "Upper Disturbed Zone" and helps to establish correlations into sections in the southern West Canada and eastern Mohawk Valleys. To the southeast, the coarse-grained intervals of Trenton Falls grade into substantially thinner and finer-grained, condensed shaly nodular facies before developing into siliciclastic dominated shale successions of the upper Flat Creek.

Prospect Quarry Member

The Prospect Quarry member at Trenton Falls is easily delineated lithologically and by marker horizons. This succession, named for exposures in the Prospect Quarry on the west bank of West Canada Creek just south of the village of Prospect, is defined as another shallowing-upward succession that is capped by the overlying Steuben Formation. The lower portion of this shallowing-upward succession is designated the Prospect Quarry Member, and is ~ 3m thick. It is substantially coarser overall than the subjacent Spillway Member, yet is distinguished from the overlying Steuben Formation in that it still contains a good proportion of finer-grained limestones and interbedded shales. The Prospect Quarry Member commences immediately above the phosphatic capping bed of the Spillway Member with a thin (.5 m) interval of interbedded shales and tabular calcisiltites. These grade upward into thin to medium bedded coquinal packstones and grainstones. Near the top, the beds become less shaly and show evidence of increased bed amalgamation and allochems show a higher degree of abrasion.

The Prospect Quarry Member is recognized at Trenton Falls by its basal and capping beds. The "Upper Disturbed Zone" and phosphatic capping bed of the Spillway Member are sharply overlain by much finer-grained calcisiltites and shales. This change represents yet another flooding style contact. The upper contact with the overlying Steuben Formation is marked by the rapid transition into very coarse-grained, massive bedded, crinoid and brachiopod coquinal grainstones. Although an upward-extension of the same shallowing-upward succession, the Steuben is set off as the shallowest facies of the entire Trenton Falls succession.

Trenton Group Formations: Steuben Formation

The Steuben Formation represents the uppermost unit exposed in the Trenton Group in the Mohawk Valley; its top contact is represented by an unconformity in this region. To the northwest, the Steuben transitions upward into the overlying Hillier Formation. It was applied to the upper unit of the Cobourg (Kay, 1943). This unit is exceptionally coarse-grained and lacking in shaly interbeds characteristic of the Trenton Group. It has been recognized as the distinct "crystalline beds" since the very first geological surveys. The Steuben Formation takes its name from the exposures on Steuben Creek, just west of Barneveld, New York (just west of Trenton Falls; **figure 24**).

The Steuben Formation is represented by massively bedded, very coarse-grained, skeletal grainstones and coquinas. These commonly show evidence for cross-bedding and sediment reworking. In the Trenton Falls region the Steuben is represented by two main coarse-grained intervals interrupted by a thin interval of shaly, recessive weathering, wavy-bedded wackestones

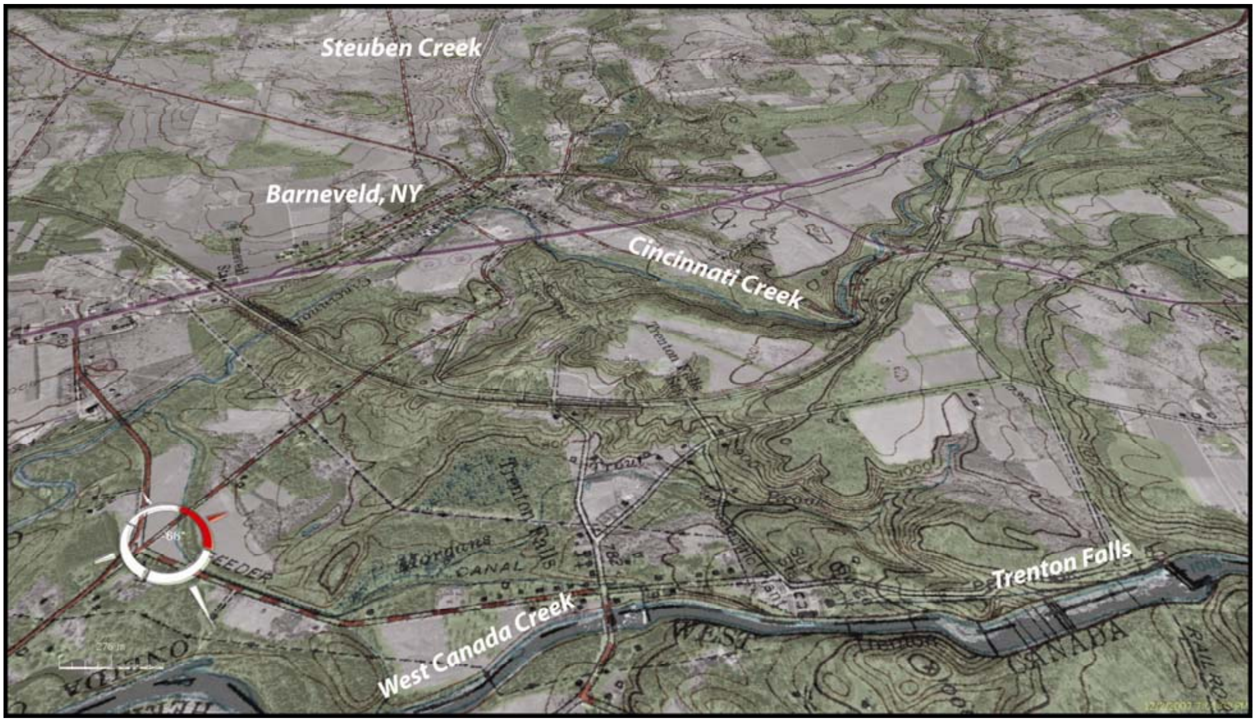


Figure 24: Type region for the Steuben Formation in the Steuben Creek a tributary to Cincinnati Creek and West Canada Creek. Map produced using images obtained from NASA World Wind digital globe software with overlays of 7.5' topographic map superimposed on 1m USGS digital orthophoto. View is to the northwest.

and packstones. Thus delineated, the Steuben is separated into two main units separated by the middle shaly interval. Due to its association with the underlying Prospect Quarry Member of the Rust Formation, the lower member of the Steuben appears as the upper half of a shallowing or coarsening-upward succession. The upper member, referred to herein as the Remsen Falls member displays a similar upward-coarsening pattern from the base of the middle shaly interval. The Steuben Formation at Trenton Falls is approximately eleven meters (30') thick, although it is difficult to assess its total thickness given that the top of the gorge is composed of Steuben and the uppermost contact is not observed in this location. However to the north near Remsen on Cincinnati Creek, the total thickness of the entire Steuben is measured at nearly twenty four meters of section.

Lower Member

The informal "lower Member" of the Steuben represents the continuation of the upward-shallowing pattern initiated in the Prospect Quarry Member of the underlying Rust Formation (**figure 25**). The lower member is defined as the succession of coarse-grained grainstones

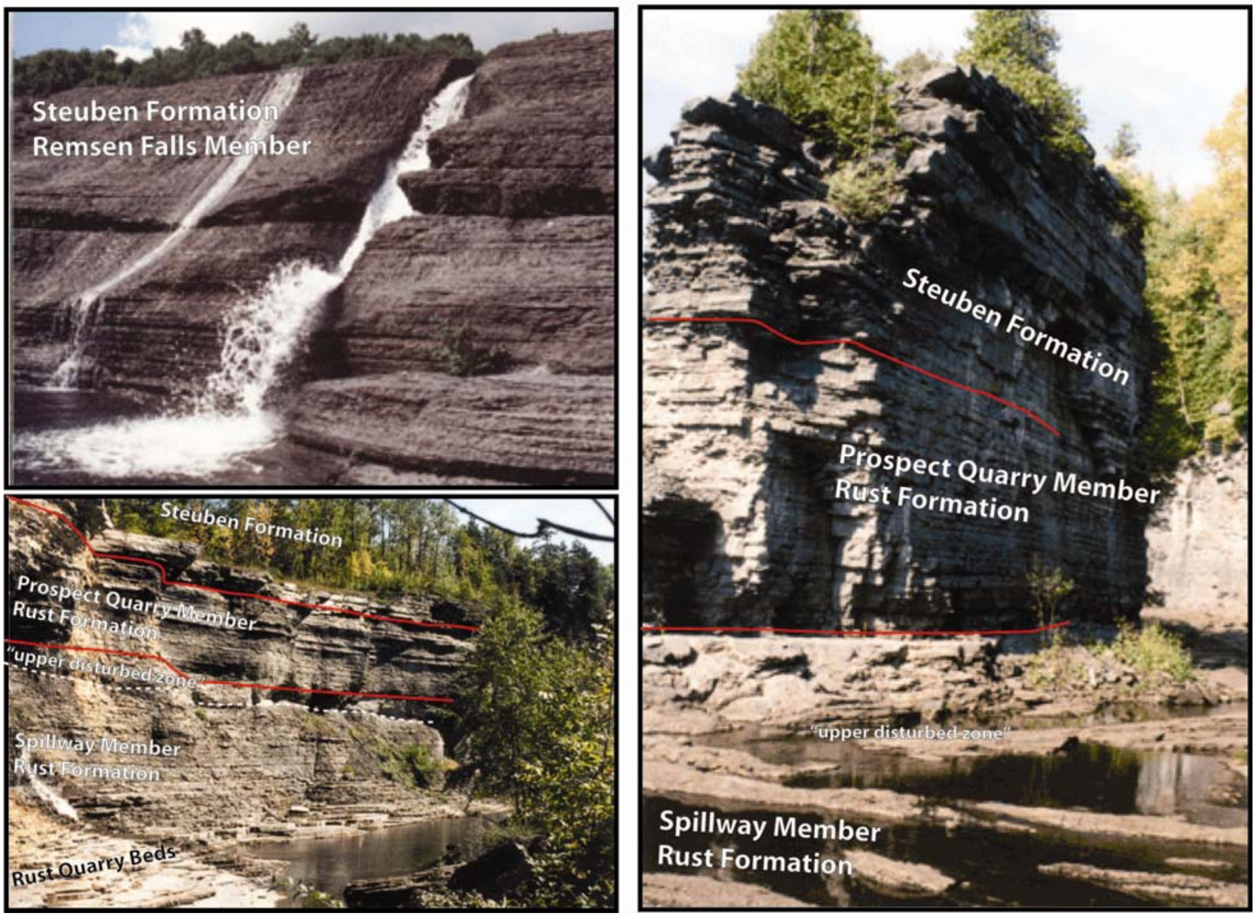


Figure 25: Outcrop photographs showing the upper Rust Formation through the Steuben Formation and associated sub-units within the West Canada Creek gorge between Trenton Falls and Remsen Falls, New York.

present above the Prospect Quarry to the base of the middle shaly wackestones and packstones at the base of the upper member. Within this interval, the lower Steuben member is essentially a homogeneous cross-bedded coarse crinoid and brachiopod grainstone.

The basal contact of the lower member is thought to be conformable with the underlying Rust at Trenton Falls. The disappearance of shaly interbeds and the increased welding of massive grainstone beds with one another is the lithologic signature that records the basal Steuben. Although the lower Member is essentially a homogeneous succession, the upper meter or so of the member shows increased rusty mineral staining with minor evidence of hard- or at least firm-ground development. This interval appears to represent the transition to a more sediment starved condition, just prior to the deposition of the middle shaly interval.

"Remsen Falls" Member

From the base of the middle shaly, wacke- to packstone interval, the remainder of the Steuben Formation grades upward in one fairly continuous shallowing and coarsening upward succession. It is very similar in motif to the subjacent shallowing-upward succession (Prospect Quarry to Lower Steuben), but it is relatively thin and nearly equal in thickness to the Lower Steuben by itself. The Upper Steuben at Remsen Falls is composed of approximately six meters of thin to medium bedded shaly bioclastic limestones. These grade upward into coarse-grained, massively bedded, crinoidal limestones similar to the lower member.

Given the discrete lithologic texture of the Steuben, this interval is readily discerned and correlatable as a unit. The sharp transition from the massive coarse grainstone facies of the lower member into the shalier middle Steuben represents a substantial flooding horizon. Similarly, the upper contact of the Steuben Formation is represented by an even more substantial flooding surface that marks the end of carbonate deposition in the Trenton Falls area and the ultimate demise of the GACB in this region. Superjacent to this contact are the dark siliciclastic shales of the Indian Castle and siltstones of the Frankfort Formation over the top of the carbonate dominated Trenton Group. The Steuben does not have many internal marker horizons used in correlation to date.

Nonetheless, there are several reentrants in the otherwise strongly bedded Steuben in the falls at Remsen. These reentrants, the first near the water line at the top of the plunge pool and the other approximately 2.5 meters higher, occasionally show swelling clays. They often support vegetation growth in the face of the falls as is common with other known K-bentonite horizons in the region. In several sections to the northeast including near Watertown there are several reentrants in similar positions. It is suspected, but not confirmed that these notches might contain the equivalents of a pair of bentonites studied by Mitchell and colleagues (1994) from the uppermost Dolgeville Formation in the East Canada Creek Valley. The Manheim and overlying West Crum K-bentonites are found below the top (2.5 m and 1.0 m respectively) of the Dolgeville. Given the detailed correlations by Baird and Brett (2002) and previous authors for the Upper Trenton units, this correlation is feasible however further it needs further investigation.

In terms of biostratigraphic markers, there are few occurrences of distinctive taxa that are used elsewhere (i.e. in the Dolgeville) to help in correlations. In some places, the top of the Remsen Falls member shows evidence for hardgrounds with upright crinoid stems that were attached to the substrate and were buried in place by coarse skeletal debris. Although similar beds are well known and well-correlated over wide areas from the time equivalent Point Pleasant Formation from the Ohio River region near Cincinnati, OH (McLaughlin & Brett, 2007) such correlation in New York is not yet possible. Moreover, although not recognized due to the lack of graptolites in most sections, Brett & Baird (2002) suggest that the biostratigraphic boundary between the *Orthograptus ruedemani* and *Climacograptus spiniferus* graptolite zones may occur within the Steuben Formation.

Chapter 4: Stratigraphic Summary of the Ashbyan to Mohawkian Interval of Central Pennsylvania

ABSTRACT

In order to investigate the geodynamic evolution of eastern Laurentia in the last stand of the Great American Carbonate Bank, a review and an updated synthesis of the stratigraphy of rock units equivalent to the Chazy, Black River and Trenton groups of the type region is necessary. To date, no large-scale compendium of stratigraphic nomenclature exists for this interval, and given the importance of these rocks for both economic and historical research, the following discussion serves in this capacity. The following discussion focuses on the stratigraphy of central Pennsylvania and adjacent regions. Collectively, these regions represent the greatest continuous exposures of Upper Ordovician strata from the northeastern United States. Moreover, these rocks are especially important because of their geographic positions relative to the Ordovician Taconic Foreland Basin, their proximity to the type Mohawkian region, and for establishing continuity of stratigraphic section with the overlying Cincinnatian Series. This discussion is intended to establish the general lithologic and biostratigraphic characteristics of the Ashbyan through Mohawkian interval for central Pennsylvania. Moreover, its construction will also help to establish a framework from which a number of long-standing inter-regional correlations will be considered. Specifically worth considering are the stage to sub-stage level correlations for Trenton Group strata of the central Pennsylvania region. Subsequently, this synthesis will be used to provide stratigraphic constraints for a comparative analysis of depositional sequences throughout the ~ 10 million years of deposition in the Late Ordovician during the Taconic Orogeny.

INTRODUCTION

Stratigraphic Framework & Review of Previous Work

Upper Ordovician strata are exposed in a number of locations in central and south-central Pennsylvania where they form part of the Ridge and Valley province (**figure 1**; **figure 2**). These

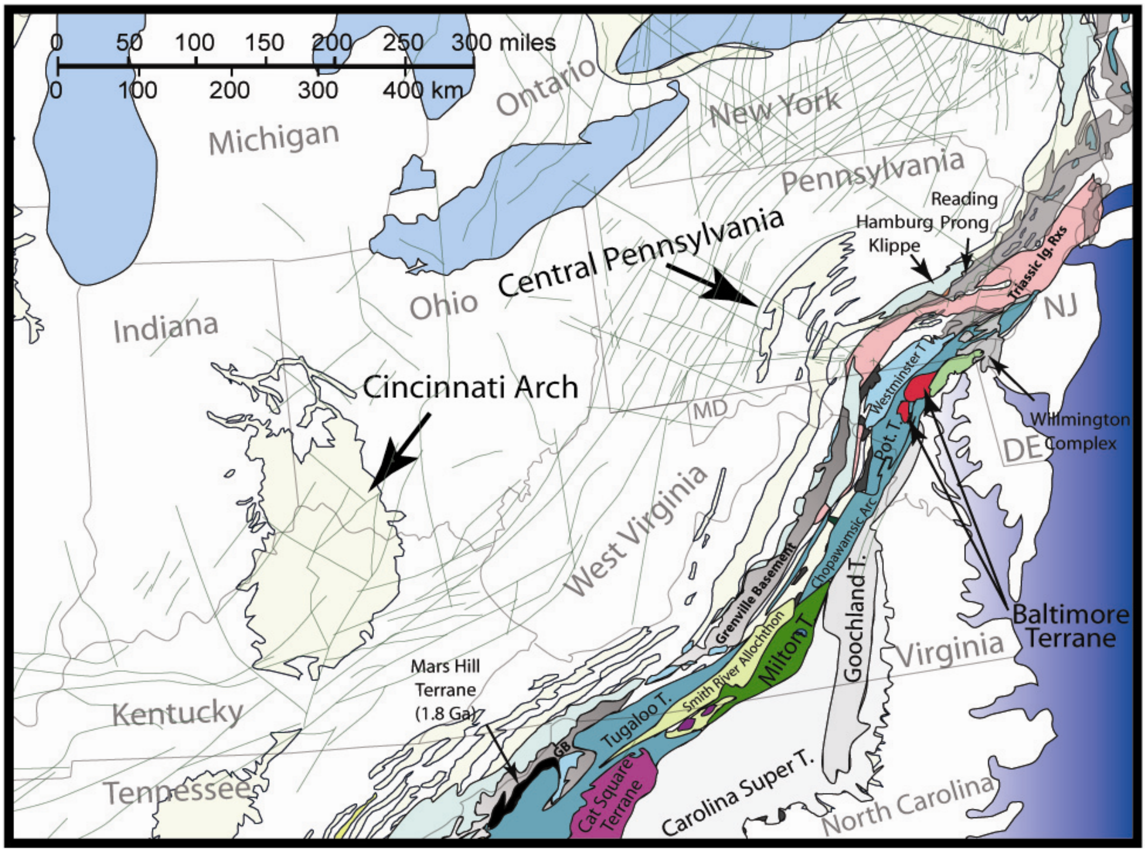


Figure 1: Ordovician outcrop areas (light yellow) of the east-central to mid-Atlantic region. Also shown are fault and lineament trends as well as major tectonic elements of the Appalachian region. These include a number of exotic terranes, allochthons, basement uplifts, and potential Ordovician volcanic arcs. Most of which appear to have been emplaced during the early to middle Paleozoic. (Abbreviations: T = Terrane, Pot. T = Poteet Terrane, GB = Grenville Basement).

rocks are also exposed in the Cumberland and Lebanon Valley sections of the Great Valley immediately northwest of allochthonous strata of the Blue Ridge Province and numerous tectonic terranes of the Piedmont region (see figure 1). Most work on these carbonate-dominated strata was initially reflective of nomenclature used in New York State although exact time equivalents of specific units have been debated. The terms Beekmantown, Chazy, Black River, and Trenton

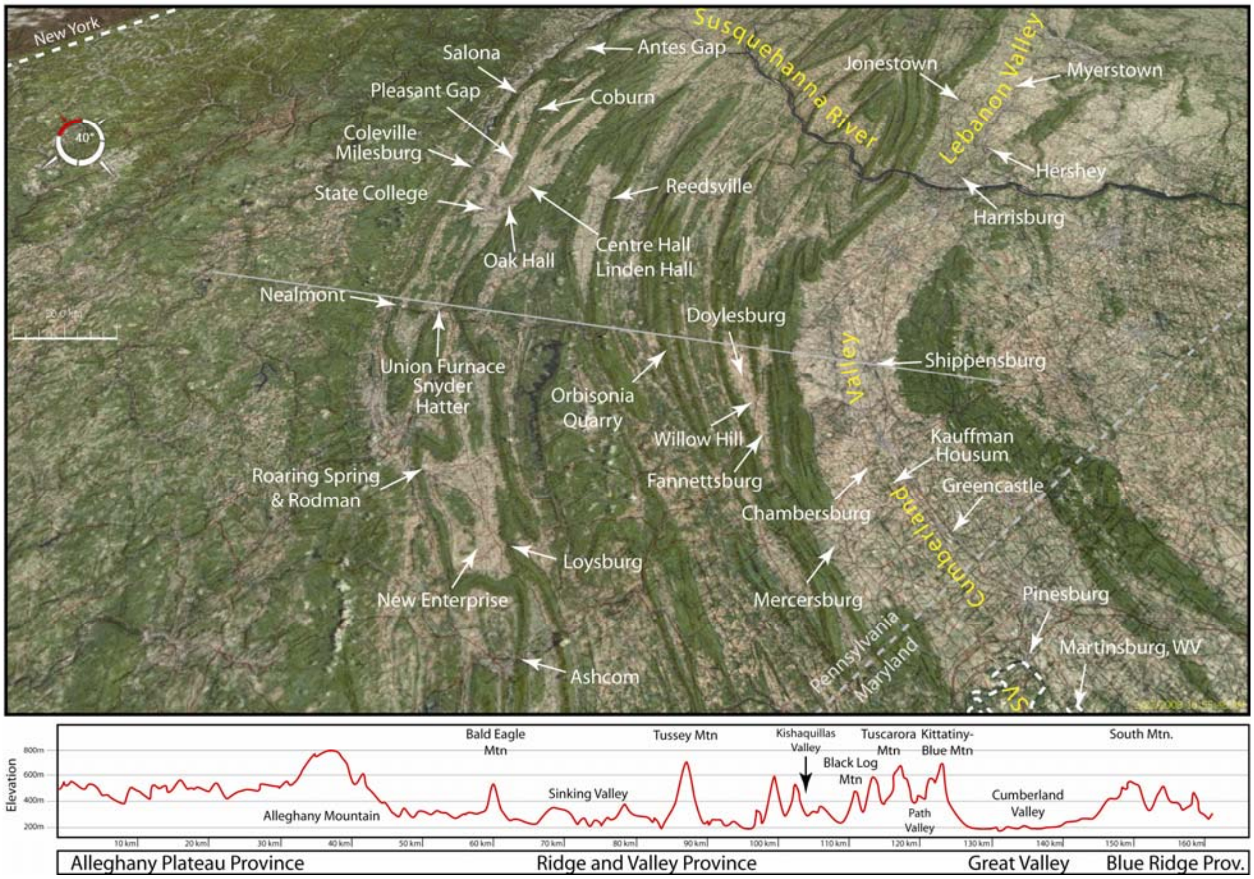


Figure 2 Type outcrop localities for stratigraphic intervals of the Upper Ordovician of the Central Pennsylvania Region and regional topographic profile showing key geographic features along the relative position of the Tyrone-Mt. Union Lineament. Perspective view map produced using images (topographic & satellite) obtained from NASA World Wind digital globe software. Vertical scale in 200 m increments, horizontal scale in 10 km increments.

have been applied in Pennsylvania owing to strong similarities between lithology and fauna from the type localities in New York. In the second geological survey report (of D'Invilliers, 1884), New York-influenced terms were used for these strata in the area of State College, Centre County. These included Trenton, Chazy, and Calciferous (as reported by Swartz et al, 1955) (**figure 3**).

Subsequently, Collie (1903) recognized nearly 183 m (603') of Trenton strata, approximately 28 m (93') of Black River strata, 77 meters (253') of "Stones River" strata, and 1464 m (4803') of Beekmantown strata (the uppermost 61 m of which is potentially correlative of basal Chazy; Wagner, 1963; Ryder et al., 2001). The stratigraphic determination was made on

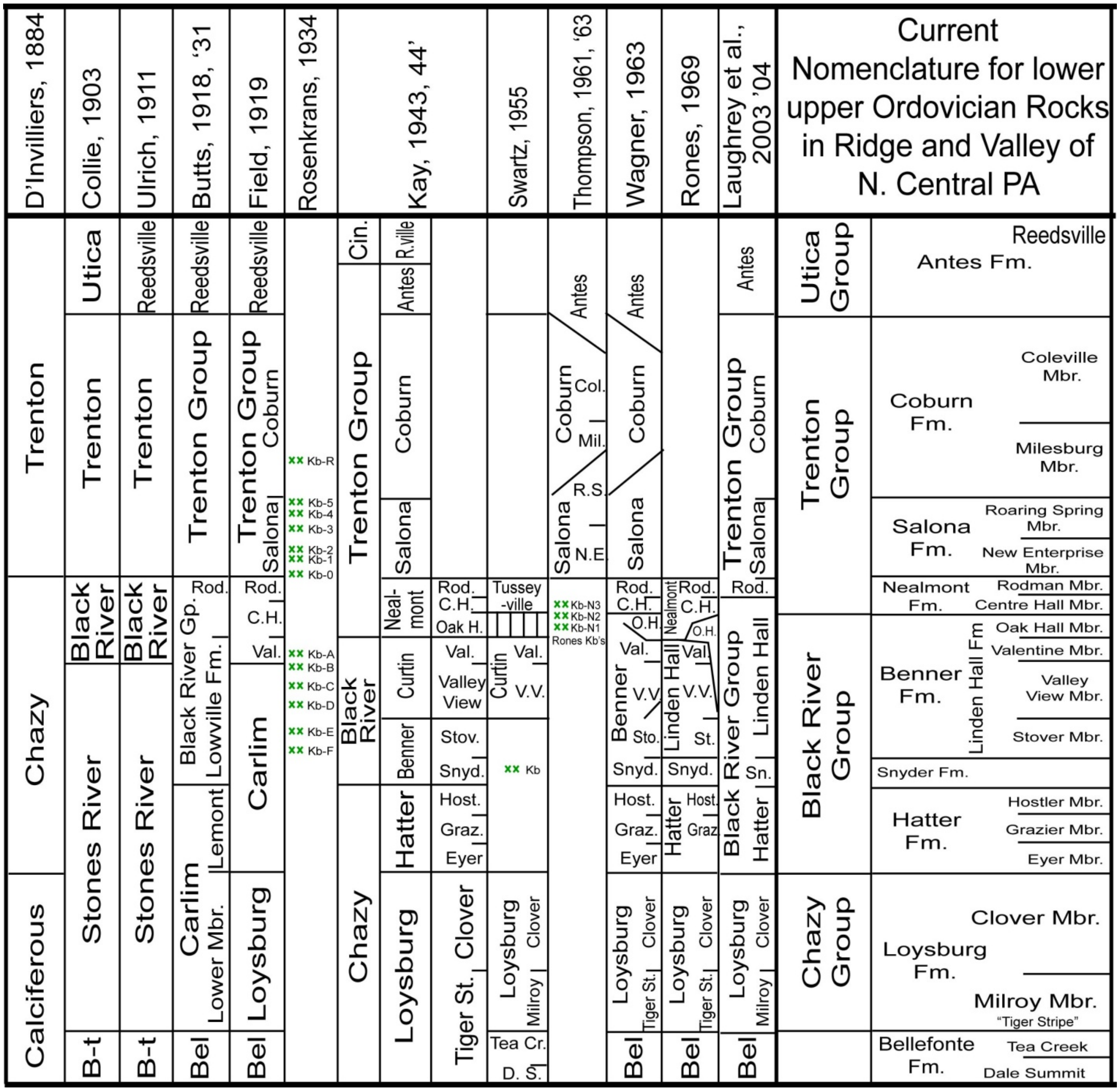


Figure 3: Historical review of lithostratigraphic classification of the Ashbyan, Turinian, Chatfieldian interval of the northwestern Valley and Ridge Province (north central Pennsylvania). Lithostratigraphic nomenclature is generally as recognized by on the Geologic Map of Pennsylvania (see Berg, 1980). The terms Benner and Linden Hall have been used interchangeably by recent authors although Benner is not used as it was originally defined by Kay (1944). Abbreviations: Kb – indicates bentonite layer, Cin., = Cincinnatian, R'ville = Reedsville, Col. = Coleville Member, Mil. = Milesburg Member, R.S. = Roaring Spring Member, N.E. = New Enterprise Member, Rod. = Rodman Limestone, C.H = Center Hall Limestone, Oak H. = Oak Hall Limestone, Val. = Valentine Member, V.V = Valley View Member, Stov. = Stover Member, Snyd. = Snyder Member, Host. = Hostler Member, Graz. = Grazier Member, B-t = Beekmantown, Bel = Bellefonte Dolostone, Tea Cr. = Tea Creek Member, D.S. = Dale Summit Sandstone

both lithologic and faunal grounds. The term Stones River was applied by Collie (1903) based on lithologic similarities to outcrops in the Tennessee and the Shenandoah Valley of Virginia. Faunally, the "Stones River" of Pennsylvania was considered an equivalent of the Chazy Group, but was a much finer-grained, lithographic limestone compared to typical Chazy. Correlations were further supported when Ulrich and Cushing (1910) subdivided the Beekmantown strata and established regional correlations from central Pennsylvania to the Great Valley region where the Stones River was located between the Chambersburg Group (equivalent to the Black River Group) and Beekmantown strata below. In total, the CBRT strata are approximately 350 meters thick in the central Pennsylvania region.

Stratigraphic boundaries of the study interval: lower contact – basal Ashbyan

As a result of these early workers, including Ulrich (1910), Butts (1918), Fields (1919), and Kay (1943, 1944), strata in central Pennsylvania were not only prescribed to New York lithostratigraphic units, but based on their faunas were also assigned within the stage nomenclature of the type region (**figure 4**). On a lithologic and faunal basis, the exact base of the CBRT interval has been an ongoing challenge and is still not firmly established in Pennsylvania. Lithologically, the dolomite or magnesium-rich interval below the "Stones River" interval has generally been equated with the Beekmantown Group while the overlying dolomitic limestones are considered to be equivalent of Chazy and Black River strata. Key faunal zones, found in some levels of the Beekmantown, enable its correlation with the type section (Ulrich, 1911) - especially below the uppermost unit (referred to as the Tea Creek Member of the Bellefonte Dolostone; Swartz et al, 1955). These intervals establish confident correlations into New York and elsewhere in Pennsylvania (Swartz et al. 1955). However, as the Tea Creek Dolostone of Pennsylvania is highly dolomitized there is little biostratigraphic evidence for establishing an exact chronology for this unit.

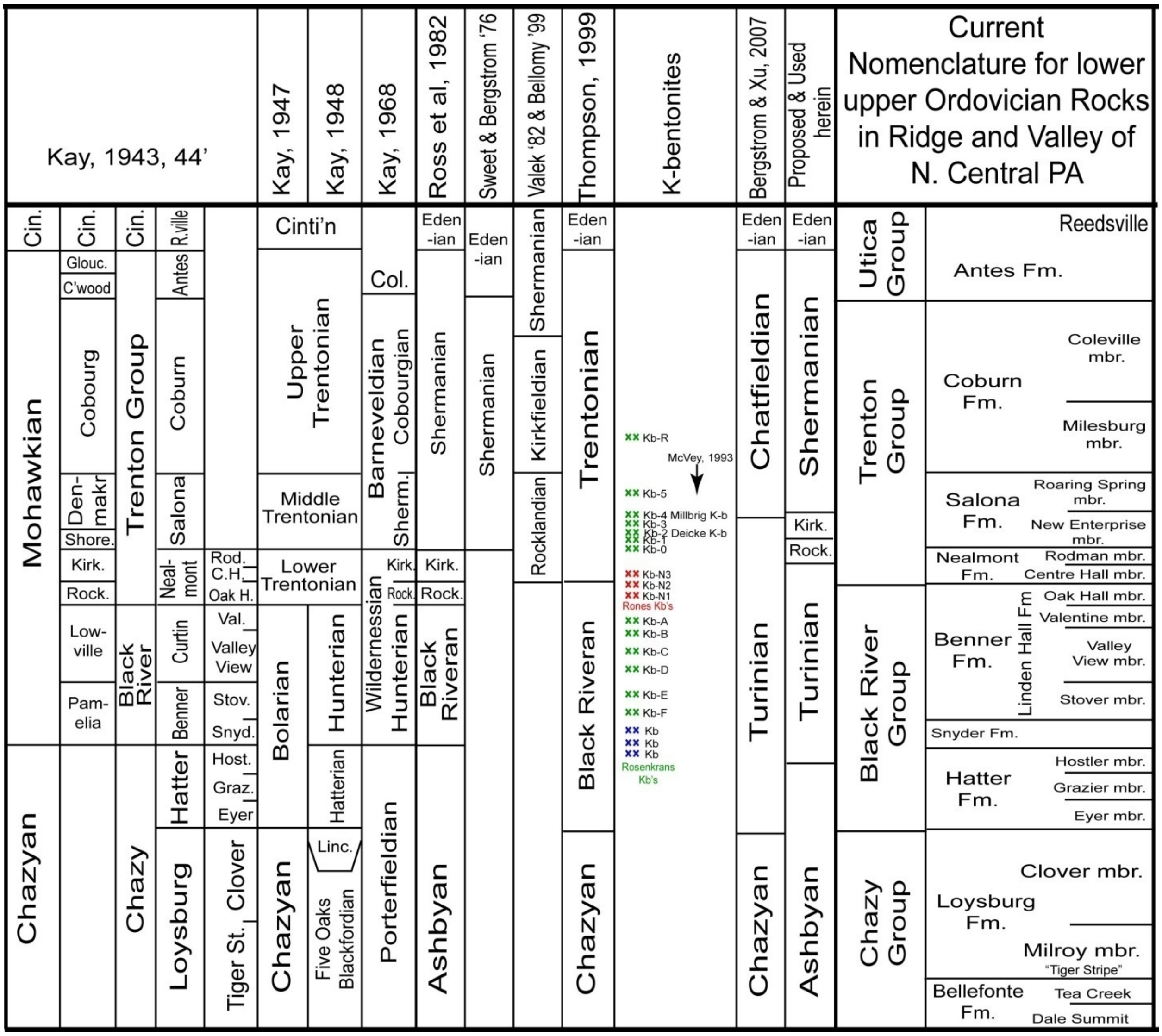


Figure 4: Review of chronostratigraphic assessments as applied to outcrops in the Ridge and Valley province and within the type region of New York State. Primary contributions were made by Kay in a number of publications. Most recent workers consider chronostratigraphic assessments to indicate "Black Riveran" to be equivalent to pre-Nealmont Formations (Benner/Linden Hall, Snyder, and Hatter Formations) while "Trentonian" strata include Nealmont, Salona, Coburn, and at least a portion of the Antes Formation). Note, Barneveldian was suggested by Fisher (1962) to replace Trentonian and is reflected in Kay's terminology of 1968.

However, as established by Swain (1957) ostracod biozones have been used to correlate the overlying Loysburg with the uppermost Day Point and Crown Point Formations of the type Chazy. This correlation leaves the basal quartz-rich and occasionally feldspathic intervals of the Day Point unrepresented in central Pennsylvania. Nonetheless, the Tea Creek dolostone, itself a distinctive red-mottled unit, contains a lithologically unique marker bed at its base called the

Dale Summit Sandstone. The Dale Summit member is a well-rounded, and frosted (?), quartz sandstone with quartz pebbles, intraformational limestone/dolostone clasts, and chert fragments (Swartz et al., 1955; Hunter, & Parizek, 1979). Thus the quartz-rich lithology is similar to the base of the Day Point Formation (Head Member) and is very similar to facies in the base of the Stone's River Group to the south that demarcate the Knox unconformity. Hence it is suggested here that the Dale Summit Sandstone sits immediately above the position of the Knox Unconformity and represents the first transgressive deposit of the Tippecanoe transgression with the Tea Creek representing the onset of carbonate deposition. This interval thus sits in the approximate position of the lower St. Peter Sandstone of the mid-continent region.

Stratigraphic boundaries of the study interval: upper contact – upper Shermanian

Ulrich (1911) named the 300 meter-thick, Reedsville Formation from the Kishacoquillas Valley of Pennsylvania and classified it faunally as representative of the Edenian to Maysvillian on the basis of a number of key fossils including the trilobite *Flexicalymene meeki*. *F. meeki* is a prominent faunal element in the Kope Formation of Ohio which is in the type Edenian interval. This age-determination was supported by Kay (1944) and is still supported today (Ryder, 1991). The lower portion (~120 m) of Ulrich's Reedsville was separated from it by Kay (1944) on the grounds that it is distinctively shalier and darker grey to black in color. It also contains a somewhat different faunal association in the lower beds dominated by the trilobite *Triarthrus*. This interval was classified the Antes Shale and separates typical Trenton lithologies and faunas from those of the Reedsville. In the Nittany Valley of Centre County and the Nippenose Valley in Clinton County (northeast of the Nittany Valley), the Antes rests sharply on typical Trenton strata (the Coburn Formation). Thus it sits in the stratigraphic position of black shales above the type-Trenton. Hence, Kay (1944) considered these shales to be equivalent of the Collingwood

Shale of Ontario and the upper Utica (Holland Patent) Shale of the western Mohawk Valley of New York.

In the northwestern Ridge and Valley of Pennsylvania, graptolite faunas indicate a *Climacograptus spiniferus* to *Geniculograptus pygmaeus* biozone (Edenian age) for the Antes Shale in this region (Lehman et al., 2002). Thus at least in the western segments of the Valley and Ridge – the base of the Antes Shale is coeval with the Holland Patent or upper Indian Castle shale of the type Mohawkian and is time equivalent with the Fulton Shales at the base of the Kope Formation of the Cincinnatian of Ohio. As this correlation has many biostratigraphic and lithostratigraphic foundations, recognition of this interval helps form a strong correlative tie line between these two type areas. As suggested by Lehman et al., 2002, the base of the Edenian, and by association the top of the Shermanian, is established at the basal contact of the Antes with the underlying Coburn Formation where they suggest a possible unconformity similar to New York is developed.

<u>Internal Stratigraphic Boundaries: Stage –Substage boundaries</u>

With numerous intervening strata containing characteristic Chazy, Black River and Trenton lithologies and faunas, it has been generally accepted that these strata represent chronostratigraphic equivalents of the type interval (as shown in **figure 3**). Nonetheless some distinctive lithologic and biostratigraphic differences, especially in the Trenton, do occur in central Pennsylvania that present some difficulty for establishing exact correlations with typical stage and sub-stage classifications of the type region.

Ashbyan Stage

The lowest chronostratigraphic unit, the Chazyan or Ashbyan Stage (now preferred), has been recognized imperfectly since the earliest surveys. Biostratigraphically, the base of the Upper Ordovician and the Ashbyan is now defined at the base of the *Nemagraptus gracilis* graptolite zone which corresponds to a position in the middle of both the *Pygodus anserinus* and the *Cahabagnathus sweeti* conodont zones (Webby et al., 2004; **figure 5**). Given the lack of

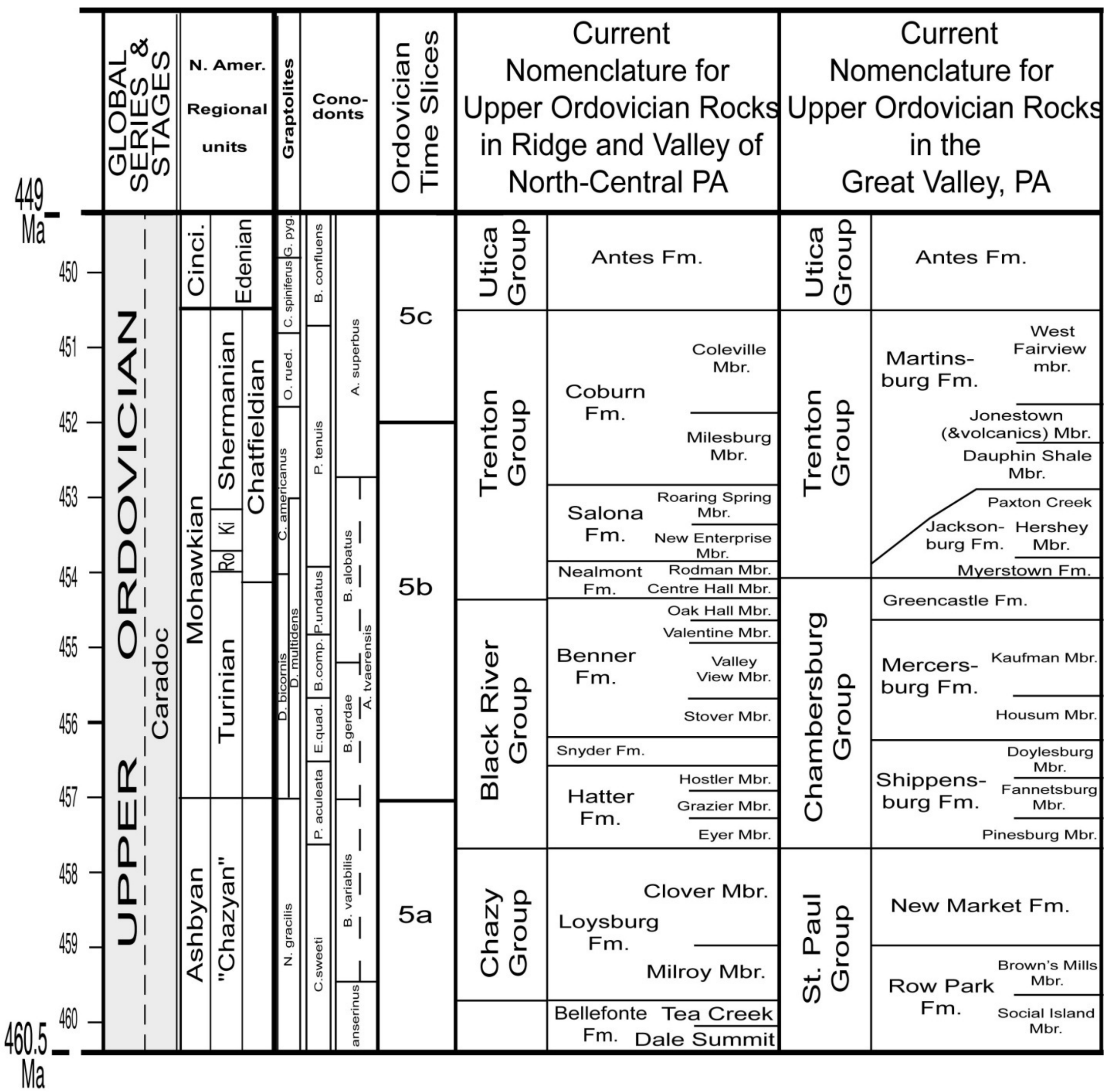


Figure 5: Biostratigraphy and lithostratigraphic nomenclature, as currently applied for the Ridge and Valley Province of north central Pennsylvania and the Cumberland to Lebanon Valley regions of the Great Valley in south central

Pennsylvania. The base of the rock succession as applied herein is established at the approximate position of the Knox Unconformity that separates the uppermost members of the Bellefonte Dolomite Formation from the lower Bellefonte. The position of the unconformity is established at the base of the Dale Summit Sandstone member (locally developed into a quartz-chert-limestone conglomerate) of the Bellefonte. The unconformity thus caps the remainder of the Beekmantown strata.

graptolites and identified conodonts from the upper Bellefonte through Loysburg interval, the base of the Ashbyan is established based on biostratigraphic correlation of superjacent and subjacent macrofaunal zones as well as the approximate position of the Knox Unconformity that forms the base of the type Chazy (Butts & Moore, 1936; Kay, 1944; Ryder et al., 2001). In central Pennsylvania this horizon sits below the top of the Bellefonte Dolostone as discussed above.

The upper Ashbyan chronozone boundary is now established at the base of the *Diplograptus bicornis* graptolite zone and coincides with the base of the *Baltoniodus gerdae* conodont subzone of the longer ranging *Amorphognathus tvaerensis* conodont zone (Webby et al, 2004). It also occurs near the top of the *Plectodina aculeata* mid-continent conodont zone. In central Pennsylvania, conodont and graptolite assessments are insufficient to produce any meaningful correlations for this interval. However, based on correlation of other macrofaunal elements including the studies of ostracod biozones as established by Swain (1957) – the Turinian-Ashbyan contact may lie within the interval between the Hatter Formation and the overlying Snyder Formation (see **figure 4**: Ryder et al., 2001). This is generally consistent with most previous stratigraphic assessments, except as applied by Thompson (1999) where the base of the Black Riveran (or Turinian) was drawn to include the Hatter Formation. Nonetheless, the Eyer to basal Grazier members of the Hatter in Pennsylvania are dominated by facies that are reminiscent of the Beech Member of the Crown Point Formation of the type Chazy as described by Oxley & Kay (1959; **figure 6**). In contrast, the overlying Hostler Member of the Hatter formation is a siliceous interval, although mostly a limestone it typically weathers buff and has

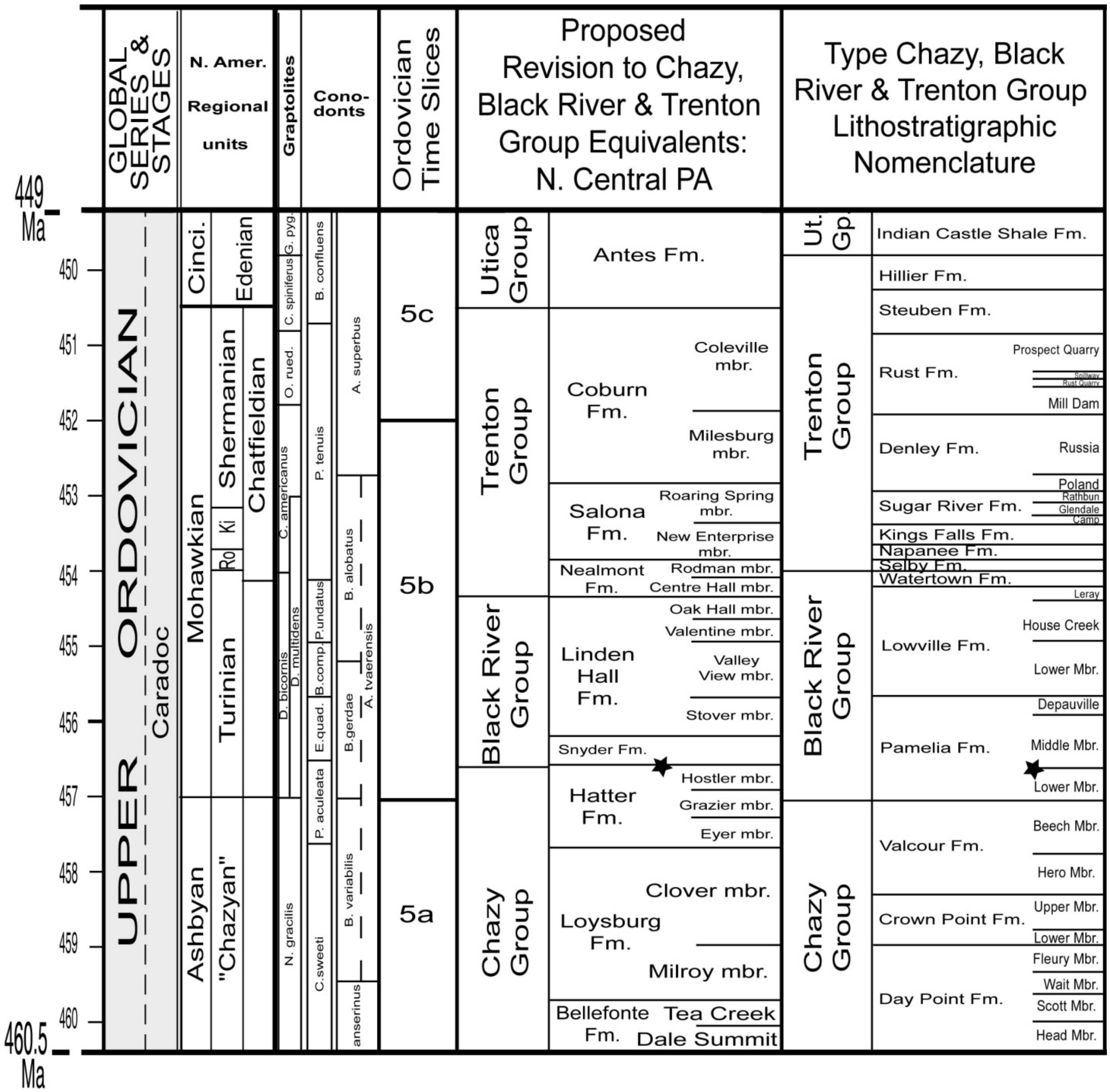


Figure 6: Lithostratigraphic nomenclature and suggested correlations between central Pennsylvania and the type region of the Chazy, Black River, and Trenton groups of the New York State region. Star denotes the position of the Chazy-Black River Unconformity as recognized herein.

facies similar to the Loysburg Formation and to the Pamelia Formation of the type region. Hence, as suggested herein and shown in **figure 6**, the Ashbyan-Turinian boundary may actually correspond to a position within the Hatter Formation and may lie within the Grazier Member.

Turinian Stage

As discussed elsewhere, the Turinian Stage was named for exposures in the Black River Valley of New York State and was defined originally to include all of the units within the Black River Group. The chronostratigraphic base of the Turinian Stage is drawn at the base of the Pamelia Formation (**see figure 6**). In the Black River Valley the Pamelia is unconformable on Precambrian to earliest Ordovician rocks. In the northern Champlain Valley, as recognized by Fisher (1968), the Pamelia Formation rests immediately above the Chazy Limestones with only minor evidence for disconformity. Based on Fisher's correlation the base of the Turinian Stage coincides with the top of the Chazy Limestones and lies at the base of the lowest beds of the Pamelia Formation with very little evidence for an erosive gap. In central Pennsylvania, the position of the basal Turinian stage is synonymous with the top of the Ashbyan stage as previously discussed. Biostratigraphically, the base of the Turinian is roughly coincident with the base of the *Diplograptus bicornis* graptolite zone and the base of the *Baltoniodus gerdae* conodont subzone as noted previously.

In the type region, the top of the Turinian Stage was established at the top of the Black River Group and thus included the Pamelia, Lowville, and Watertown Limestones (Walker, 1973; Fisher 1977), the latter of which was the Black River Limestone sensu early workers. This stratigraphic horizon is roughly coincident with the top of the *Phragmodus undatus* (defined at the FAD of *Plectodina tenuis*) conodont biozone and falls near the top of the *Diplograptus bicornis* graptolite zone. The recognition of *Plectodina tenuis* conodonts in the uppermost Black River Group (Chaumont Formation) in Ontario by Barnes (1967; reconfirmed pers. comm., 2008) helps constrain the *P. undatus – P. tenuis* zonal boundary to the approximate position of the Black River – Trenton lithostratigraphic boundary (Titus & Cameron, 1976; Cameron & Mangion, 1977). This observation is critical as it also helps establish a recognizable boundary for the top of the Turinian and the base of Kay's (1943, 1944, 1968) Rocklandian stage (see below). Nonetheless owing to the historic difficulty of correlating the chronostratigraphic units out of New York, the top of the Turinian was truncated to the position just below the Millbrig K-bentonite and the base of the Millbrig K-bentonite was proposed as the base of the Chatfieldian Stage (Leslie & Bergström, 1995). Thus although the Turinian-Rocklandian chronostratigraphic boundary of the type region is equivalent to the *P. undatus – P. tenuis* boundary, recognition of the Millbrig K-bentonite in New York (Mitchell et al, 2004), puts the top of the Turinian at the level of the Millbrig K-bentonite. Based on research of the author, this puts the top of the Turinian below the Watertown Limestone and thus places the entire Watertown Limestone in the overlying Chatfieldian Stage.

In Pennsylvania, under the proposal of Leslie & Bergström (1995), the top of the Turinian is drawn at the position of the Salona 4 K-bentonite which was proposed to be the Millbrig K-bentonite by McVey (1993) (**figure 7**). Despite a number of objections raised to this correlation as outlined herein, the loosely defined *P. undatus – P. tenuis* conodont biozone boundary has been drawn just above the Salona 4 K-bentonite (of Rosenkrans, 1934) and is thus permissive of the Millbrig correlation (Sweet, 1984; Kolata et al., 1996). The first recognized occurrence of *P. tenuis* is in a thinner-bedded shalier portion of the New Enterprise Member of the Salona Formation (Leslie, 2000) although the zone lacks the same distinct lithologic transition that often demarcates the *P. undatus – P. tenuis* boundary elsewhere (i.e. basal Lexington Group contact of Kentucky, basal Trenton Group contact of New York, basal Dolly Ridge contact of Hagan, Virginia etc.). Additional chemostratigraphic work provides some evidence suggesting that the *P. undatus-P tenuis* biozone boundary may still be place too high and that the position of the Millbrig in Pennsylvania should be re-evaluated. Recent stratigraphic

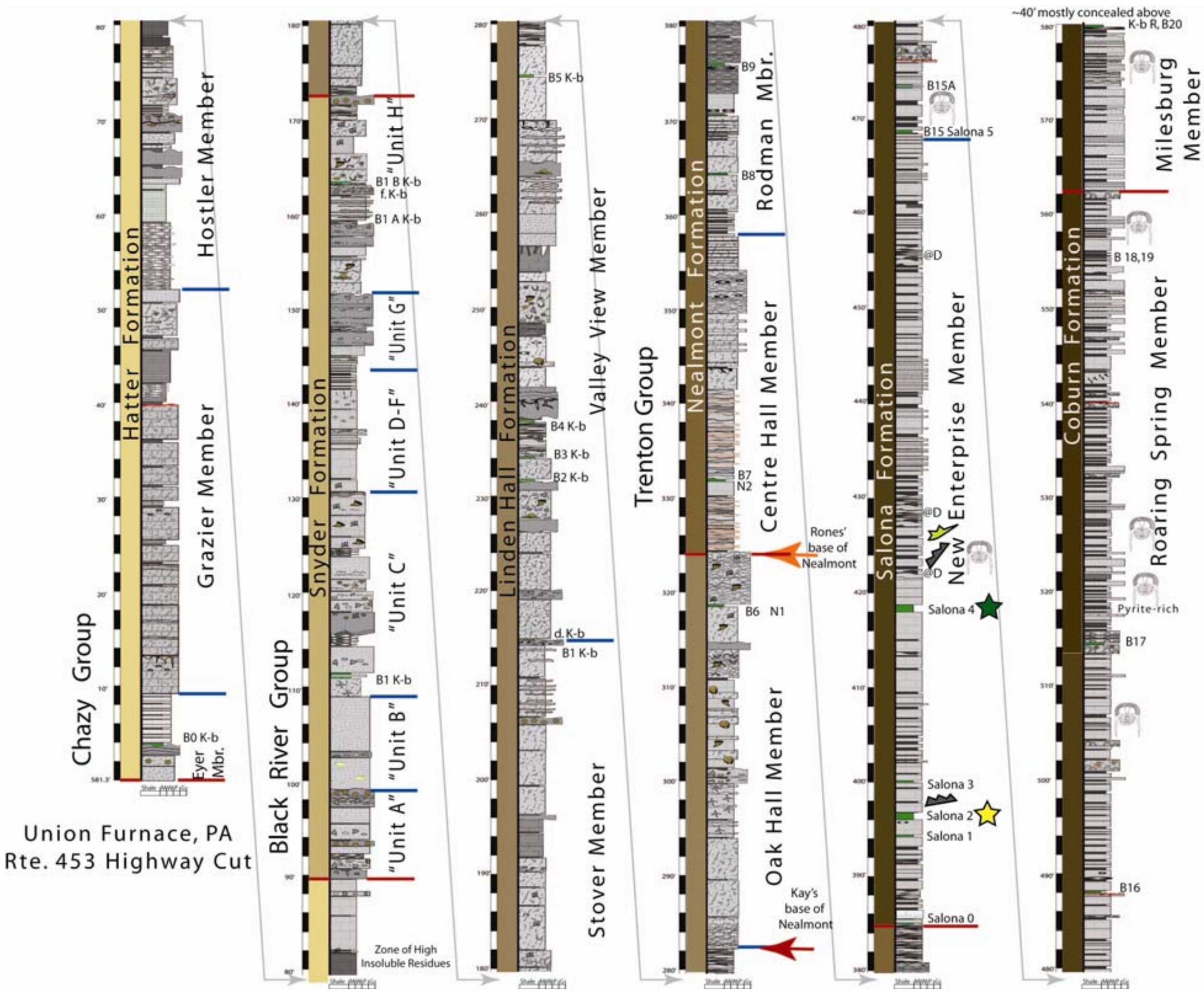


Figure 7: Stratigraphic section for the Union Furnace (Rte 453 Highway Cut) from the uppermost Loysburg Formation (Clover Member) to the lower Coburn Formation (Milesburg Member). Shown are the strata from the upper Chazy Group, the entire Black River Group, and the majority of the Trenton Group Limestones. Important to note are the positions of the base of the Nealmont as favored by Rones (1969), and Kay (1943, 1944) (red and orange arrows). Bentonites are labeled after the schemes of Rosenkrans (1932), Rones (1969), and Berkheiser & Lollis (1986). Stars indicate the controversial positions of the Deicke K-bentonite (Yellow) and the Millbrig (Green) as suggested by McVey (1993). Also shown are the positions of graptolite occurrences noted by Whitcomb (1932a, b) and used to define the C. americanus graptolite zone.

work on the Guttenberg Isotopic Carbon Excursion (GICE) (Young et al., 2005; Barta, 2004) shows that the excursion occurs above the position of the Millbrig and above the *P. undatus* – *P. tenuis* boundary in all regions of the GACB including New York, Kentucky, and Virginia. In Pennsylvania, however, the GICE as recognized by Patzkowsky and colleagues (1997) begins below the position of the inferred *P. undatus* – *P. tenuis* conodont boundary and below the position of the Salona 4 K-bentonite. If the initiation of the GICE is a synchronous event relative to the Millbrig K-bentonite as it appears to be elsewhere, the uppermost boundary of the Turinian in Pennsylvania is still not established firmly.

As suggested herein and previously by Barta (2006), available biostratigraphic, lithostratigraphic, and petrographic analysis indicate that the top Turinian boundary might occur in the middle of the Nealmont Formation as currently defined. Barta (2006) suggested that the Centre Hall Member of the Nealmont is faunally and lithologically more characteristic of the Black River, whereas the upper Rodman is faunally and lithologically characteristic of the Trenton. Collectively these data are important and suggest that Kay's (1943b, 1944, 1947, 1948) chronostratigraphic assessments of the Nealmont (as being equivalent to the Kirkfieldian) were too young (see Figure 4). In addition, the age assessments installed as a consequence of the McVey (1993), Leslie and Bergström (1995), and Leslie (2000) assessments may also be too old in that the Salona 4 K-bentonite is likely younger than the true Millbrig. Based on these presuppositions as well as the lithostratigraphic, and biostratigraphic assessments the true position of the top Turinian boundary may indeed be more aligned with the earlier stratigraphic assessments.

As originally proposed by Butts (1918), the uppermost Black River unit of central Pennsylvania was considered to be the black, massive-bedded, fossiliferous, micritic, but siliciclastic influenced, Rodman Limestone (originally defined to include both the Centre Hall and the Rodman of today's nomenclature). This unit was most similar to the original definition for the Black River Limestone in its original sense (of Cushing et al., 1910). The Rodman was further distinguished by Butts (1931) who used the term Lowville Formation for the units below the Rodman owing to their significant similarity both faunally and lithologically to the type Lowville. Thus the Nealmont Formation likely straddles the Black River – Trenton lithostratigraphic boundary and also contains the Turinian-Chatfieldian boundary (pending recognition of a more suitable Millbrig in the Nealmont). Furthermore, the Nealmont likely contains the *P. undatus – P. tenuis* zone boundary as shown in Figure 6. On these grounds, and as suggested by the appearance of the GICE in New York and Pennsylvania (Barta 2004), the lower Nealmont is likely Turinian and the upper Nealmont is likely equivalent to the Watertown-Selby interval and is thus Rocklandian or Chatfieldian in the scheme of Leslie and Bergström (1995) if the position of the Millbrig can be elucidated.

Chatfieldian Stage

In central Pennsylvania the recognition of the Chatfieldian Stage is herein considered problematic on grounds highlighted above. Despite recent investigations by Carey (2007), samples from the Salona K-bentonites at Union Furnace, Pennsylvania failed to yield apatite samples for ICP-MS analysis as used elsewhere for identification of the Millbrig and other K-bentonites (i.e. Samson et al., 1988; Emerson et al., 2004, Mitchell et al., 2004). The inability to produce a heavy mineral separate with apatite phenocrysts for chemical analysis from the Salona K-bentonites (Salona 2 and 4) prevents corroboration of McVey's (1993) correlations of these specific K-bentonites with the Deicke and the Millbrig which was originally based on analysis of whole-rock mineralogy and discrimination of bulk mineral chemistries (McVey & Huff, 1993). Although apatite phenocrysts are typically a small component of most K-bentonites they are typically present in sufficient numbers, especially in the Millbrig and Deicke K-bentonites, to allow their separation and analysis as has been done elsewhere including in nearby West Virginia, Virginia, and New York (Mitchell, 2004; Carey, 2006). The analysis of a specific mineral grain (such as apatite) is a more robust fingerprinting technique that avoids the inherent challenges posed by bulk-mineral analysis. Nonetheless, until this controversy is resolved and a suitable Millbrig K-bentonite candidate is confirmed, herein it is suggested that stage-level chronostratigraphic assessments follow those of the type-region including application of Kay's Rocklandian, Kirkfieldian, and Shermanian stages. Albeit, the application of these chronostratigraphic terms as recognized herein is not consistent as first applied in Pennsylvania (Kay, 1943, 1944).

Rocklandian, Kirkfieldian, and Shermanian Stages

As discussed in the context of the top Turinian contact above, the base of Kay's New York based Rocklandian stage can now be generally associated with the *P. undatus – P. tenuis* conodont biozone boundary. Also in the type region, the approximate base of the *Corynoides americanus* graptolite zone is projected to the approximate base of the Rocklandian (Brett & Baird, 2002; Goldman, et al., 2002). The recognition of the approximate position of the boundary helps strengthen the chronostratigraphy established by Kay (1943, 1944). In Pennsylvania, graptolites including *Corynoides americanus* (var. *calicularis*) have long been recognized in the lower Martinsburg Shale in the Cumberland Valley region immediately above the Chambersburg Limestone (Bassler, 1919). These forms have also been recognized above the Jacksonburg Limestone in the Bushkill Shale Member of the Martinsburg in more eastern portions of the Great Valley (Epstein & Berry, 1973; Parris & Cruikshank, 1992; Parris et al., 2001).

In the Ridge and Valley region, the appearance of the type species *Corynoides americanus* has not been established although graptolite occurrences in the middle of the New Enterprise and Roaring Spring members of the Salona Formation at Bedford have been noted and provide an opportunity for future investigation. Also reported by Whitcomb (1932 a,b) in the type section of the Salona, beds immediately below the Salona 3 K-bentonite (and below the suspect Millbrig K-bentonite) were documented to contain identifiable graptolite forms identified by Ruedemann as *Diplograptus amplexicaulis* (abundant) and a single specimen described as *Diplograptus (Mesograptus) mohawkensis.* As per Goldman and colleagues (1994) these are now considered to be *Rectograptus amplexicaulis* and *Normalograptus mohawkensis* respectively. An additional graptolite was recognized in a thin zone above the Salona 4 K-bentonite and was identified as *Climacograptus strictus* (now *Normalograptus brevis-strictus* sensu Goldman et al., 1994). Collectively these forms were characterized originally by Ruedemann as equivalent of the Canajoharie Shale (Whitcomb, 1934) and are representative today of the Glens Falls Limestone of New York (Goldman et al., 1994). Thus these graptolites are recognized as representing the lower *Corynoides americanus* graptolite biozone (Goldman et al., 1994). With recognition of *C. americanus* zone faunas in roughly the same interval as *P. tenuis* support the idea that the Salona Formation could range from Rocklandian to Kirkfieldian. Again the close proximity of these forms to the Salona 4 K-bentonite suggests that the Salona 4 K-bentonite is not the Millbrig which should occur lower.

In addition to these specific biostratigraphic indicators, application of chronostratigraphic zones herein is focused on re-interpreting previous litho- and biostratigraphic assessments in the context of other recent data. Field (1919) and Kay (1943, 1944) subdivided the lithostratigraphic units recognized by Butts (1918) and installed a number of stratigraphic units to reflect local sedimentologic conditions as well as distinct faunal zones. Kay used classifications in the Pennsylvania region based on shelly fossils and attempted to remove Chazyan, Black Riveran, and Trentonian from denoting any chronostratigraphic interval (Kay, 1968). Therefore, Kay (1943, 1968) installed the New York – Ontario chronology (i.e. the Rocklandian, Kirkfieldian, Shermanian, and Cobourgian), to strata in central Pennsylvania based on several important observations and presuppositions.

First the upper Rodman Limestone was known to contain abundant echinoderm faunas including the *Echinosphaerites* and was overall dominated by echinoderm fossil debris. In some levels of the Rodman (now divided into several units) beds are composed of as much as 20-30% pelmatozoan material (Rones, 1969). This was substantially different than any underlying or overlying unit (Butts, 1918, Field 1919) and was characteristic of the Kirkfield Limestone of Ontario. Second the base of the Rodman (now the Centre Hall Member), contained *Maclurites logani*, a characteristic fossil of the uppermost Black River to lowest Rocklandian of New York and Ontario. Third, Kay (1943, 1944) also identified a pronounced unconformity at the base of his Nealmont Formation (Oak Hall member) that appeared to truncate a number of underlying units. Swartz and colleagues (1955) also recognized an unconformity at the base of their Tusseyville Limestone which was another local name for Kay's Nealmont. Kay assumed this unconformity to be the Black River - Trenton unconformity of New York situated at the base of the Rocklandian. Fourth, the Rodman interval showed a significant increase in the amount of quartz silt and interbedded argillaceous sediment which was characteristic of the Trenton of New York. Kay also recognized the interbedded, turbidite-style deposition in the Salona and implied that it was most similar to bedding patterns of the lower and middle Trenton Group. Fifth, Whitcomb (1932) and Kay (1943, 1944) recognized the initial incursion of *Cryptolithus tesselatus* just above the base of the Salona and these persisted upward to the base of the upper member of the Salona before becoming significantly less abundant. These trilobites then reappeared and persisted in abundance in the upper Coburn member - a pattern that was later supported by data presented by Thompson (1963). Faunally, a number of other forms including bellerophonts (*Sinuites cancellatus*), and the nautiloids (*Geisonoceras tenuistriatum,* and *Trocholites ammonius*), became abundant in the uppermost Salona. Sixth, as established by Whitcomb (1932) and Rosenkrans (1934), the Salona K-bentonites had already been equated with K-bentonites in the Martinsburg Shale in northern Virginia and in the type region at Martinsburg, West Virginia. Kay (1944) suggested that the Salona K-bentonites (at least 2 through 6) were equivalent to those found in the stratigraphic succession exposed in the interval between Sherman Falls and the lower High Falls in the Trenton Gorge at West Canada Creek. These K-bentonites are now referred to as the Sherman Falls I & II, and the Kuyahoora I & II K-bentonites (Brett et al., 2002).

Kay (1943, 1944) used this information, including the occurrence of the unconformity, *M. logani,* and lithologic similarities to suggest the position of the Rocklandian Stage. He believed the Oak Hall through the lower Centre Hall Member constituted the time equivalents of the Selby and Napanee of the type region. Likewise, he felt the Kirkfieldian Stage was represented by the *Echinosphaerites* zone of the Rodman in Pennsylvania. The incursion of *Cryptolithus* and other faunas typical of the “Canajoharie Shale” in the Salona appeared to be indicative of his Shorehamian sub-stage as defined in northeastern New York (Kay, 1968). Kay also considered the upper Salona (the Roaring Spring member) as the Denmarkian sub-stage equivalent. These were later combined to form the Shermanian Stage in subsequent chronostratigraphic models (Kay, 1968). In this scenario, the presence of the K-bentonites, the lithologic data, and the macrofaunal occurrences were viewed as significant evidence that suggested the Salona was entirely of Shermanian age and equivalent to strata up to the level of the Rust Formation of the type region.

Finally, the last stage of the Trenton recognized by Kay included the Cobourgian (Kay, 1943, 1944, 1968). In Pennsylvania, Kay recognized the distinct coarsening of facies in the upper half of the Trenton Group (Field's Coburn Formation; 1919), as well as the incursion of a number of taxa not present in the underlying Salona. Kay considered these to be significant and similar to the pattern observed in central New York in the Rust, Steuben, and their equivalents. This in fact was supported by previous investigations of Raymond (1916). With the distinct and sharp contact of the Antes at the top of the Coburn to constrain the interval, Kay and other workers considered this to be the top of the Cobourgian Stage and thus nearly identical in age to the type region in Ontario. Furthermore data reported by Lehman and colleagues (2002) suggests that graptolites in the base of the Antes Shale immediately above the Coburn were representative of the *Climacograptus spiniferus* zone and unpublished K-bentonite correlations of the lower Antes are suggested to be equivalents of the Manheim and East Canada Creek K-bentonites of New York, which also occur in the *C. spiniferus* zone (Bellomy, 1999). Thus, the Coburn of Pennsylvania has been considered to be older than Edenian.

Re-investigation of chronostratigraphic zones in Pennsylvania

Although Kay's work was progressive, some of his assumptions have been challenged. Numerous workers recognized the difficulty in using macrofauna and lithostratigraphic assessments as a basis for establishing chronostratigraphic zonation – especially as in most cases, many taxa are facies controlled. Moreover, developing models for the Taconic Orogeny, suggested that many of Kay's biostratigraphic based chronozones were actually time-transgressive or diachronous even in the type region given the impact of the orogeny (Fisher, 1962). Therefore without solid evidence for correlation of the isochronous K-bentonite seams recognized by Kay, he lacked substantial evidence for his assessments. Given the substantial regional complexity of Kay's standard nomenclature many workers simply refer to these strata as Trentonian (see Thompson, 1999 for example) to avoid confusion.

Nonetheless, few attempts have been successful in re-evaluating these chronologies using independent corroborating techniques. For instance, Valek (1982) and Bellamy (1999) investigated conodonts from the Salona and Coburn Formations which although, depauperate and often poorly preserved, suggested these units were mostly Rocklandian and Kirkfieldian to the base of the Antes on the basis that the conodont *Polyplacognathus ramosus* both at Bellefonte and at Reedsville indicated the rocks to be no younger than the *Amorphognathus tvaerensis conodont* zone. Thus this work indicated that the Coburn-Antes contact (at Reedsville, PA) recorded a significant hiatus for the total duration of much of the *C. americanus* and the entire *O. ruedemani* graptolite zones (the upper portion of the *P. tenuis* and lower *A. superbus* conodont chronozones). This was in direct contrast to previous correlations as outlined above. In contrast, Arens & Cuffey (1989) and Cuffey (1997) continued to use Kay's standard stages as applied in the northwestern portion of the Ridge and Valley (around the State College region). Nonetheless they show all formations as progressively older and diachronous to the southeast toward Reedsville, PA. These assessments were favored despite the well-established and chemically substantiated correlations of the Salona K-bentonites which acted as time planes across much of the region (Thompson, 1963, Lollis, 1984).

Therefore significant modification of Kay's chronostratigraphic standard did not occur in publication until K-bentonite analysis by McVey (1993), as reported by Kolata and colleagues (1996), suggested the Salona 4 K-bentonite (Mill Hall K-bentonite of Lollis, 1984) was the potential equivalent of the Millbrig K-bentonite. As the Millbrig was considered at the time to be Rocklandian in age, McVey's (1993) analysis indicated that Kay's age assessments at least for the Salona and underlying units were too young and that the Salona had to at least be Rocklandian making underlying units older and corroborating at least a portion of Valek (1982) and Bellamy's (1999) conodont-based age assessments. Moreover, graptolite data from the Salona (taxonomically updated after Whitcomb, 1932) suggested a lower *C. americanus* zone for the Salona. Nonetheless, it is now known from correlation of the Millbrig into New York, that that K-bentonite is pre-Rocklandian as it sits at the base of the Watertown Limestone, and predates the C. *americanus* zone, as well as the P. *undatus – P. tenuis* boundary. Therefore in Pennsylvania, although the Salona, at least in part, is Rocklandian in age especially in the interval immediately overlying the Salona 4 K-bentonite, the correlation of the Salona 4 K-bentonite with the Millbrig is likely incorrect. Furthermore the recognition of the Millbrig K-bentonite relative to the Guttenberg isotopic Carbon Excursion in the New York sections where the Rocklandian, Kirkfieldian, and Shermanian stages were defined helps warrant a re-evaluation of the chronostratigraphy of the Nealmont-Salona in central Pennsylvania **(figures 4 &6)**.

Herein, it is proposed that the uppermost Turinian strata be defined in Pennsylvania on the basis of the last occurrence of *M. logani* and the recurrent faunal zone in the Centre Hall Formation and the recognition of *P. undatus* conodonts in the Centre Hall and the subjacent Linden Hall. The age equivalent of the type-Rocklandian is defined to encompass the majority of the Rodman through at least a portion of the lower Salona (New Enterprise Member). The echinoderm-rich interval in the top of the Centre Hall and Rodman members of the Nealmont (originally considered Kirkfieldian by Kay), is thus latest Turinian to Rocklandian in age and equivalent to the echinoderm-rich Curdsville Limestone of Kentucky and the overlying Logana Member (see below). Combined with the lithologic data presented by Barta (2004), the epibole of *Cryptolithus* in the basal Salona, the basal Logana (of Kentucky), and the Napanee (of New

York), substantiated by the appearance of the GICE in all of these regions, and the approximate position of the *P. undatus* – *P. tenuis* conodont zonal boundary there is now substantial support for this age determination.

Age equivalency for the Kirkfieldian is less confidently established owing to the subtle lithologic changes in the remainder of the Salona Formation. Nonetheless, recognition of graptolites characteristic of the *C. americanus* graptolite zone, as well as abundant *Prasopora simulatrix* within the medial Salona suggest that the unit is mostly equivalent to what Ruedemann (1912) called the Canajoharie Shales and are now referred to as the Glens Falls Limestone and lower Flat Creek Shales (Kay, 1937, Brett & Baird, 2002). At Canajoharie, New York the lower "Canajoharie Shales" (term now abandoned) are interbedded with coarse fossiliferous limestones and lie below the interval referred to as the L. Flat Creek Shales. These beds occur just below the major epibole of *Prasopora simulatrix* bryozoans in the Sugar River-Shoreham equivalent, and below the *Baltoniodus alobatus* - *Amorphognathus superbus* conodont biozone boundary recognized in the Denley Limestone at the base of the Russia Member (Schopf, 1966). Given this distinction, the cross-bedded, coarse-grained lutite beds of the upper New Enterprise member (below the Salona 5 K-bentonite) may be the equivalent of the Kings Falls Limestone. As reported by Thompson (1963) this interval is known for producing crinoid plates, the calymenid/homalonotid (?) trilobite *Homalonotus* (*Brongniartella*) *trentonensis*, and abundant lingulid brachiopods which suggest at least some shallowing from underlying units. In addition, although the trilobite is not recognized in the Kirkfield of Ontario (it may have been an immigrant from Europe), Whittington (1965) notes that trilobites (*Platycorphe platycephalus*) from the Hermitage of Tennessee are very similar to the forms recognized as *H. trentonensis* by Whitcomb (1932) in Pennsylvania and may be synonymous. Nonetheless, the Hermitage

Limestone of Tennessee, has been classified (at least in part) as Kirkfieldian by Wahlman (1992) on the basis of monoplacopherans and bellerophontids.

In addition to faunal evidence, recognition of soft-sediment deformation horizons or "seismites" first in the New Enterprise and then in the Roaring Spring member of the Salona Formation often (containing *Prasopora* rudstones) represent the earliest recognized soft-sediment deformation structures in this interval of the Ordovician of Pennsylvania. In New York, the first evidence for seismites is found in the Napanee Limestone and again in the Kings Falls to lower Sugar River Formations both in the Mohawk Valley and then in the Black River Valley respectively. Thus combined with increased abundance of *Prasopora* event beds upward from this approximate interval, the upper Salona is likely equivalent to the Kirkfieldian through lower Shermanian as was supported by Arens & Cuffey (1989) and Cuffey (1997). Given that this interval falls within the long-ranging *P. tenuis* conodont zone and within the *C. americanus* graptolite zone, these data support the suggestion that the upper Salona may indeed be equivalent to the lower Shermanian, and the unconformity suggested by Valek (1982) and Bellomy (1999) may be of significantly less duration, as the Coburn is still to be accounted for.

Based on conodonts and graptolites, the uppermost Coburn was thought to be no younger than earliest Edenian (Sweet & Bergström, 1976; Sweet, 1984) and potentially as suggested by Valek (1982, unpublished M.S. thesis) and Bellomy (1999, unpublished senior thesis), the Coburn may have been no younger than latest Kirkfieldian to earliest Shermanian. However, these authors reported the Coburn to contain a low-diversity assemblage of non-diagnostic conodont forms, most of which cannot be reliably placed into a chronostratigraphic framework from section to section without relying on other correlation techniques. Nonetheless, a number of key biostratigraphic concerns should be raised based on observations from previous literature.

First, as suggested by Whitcomb (1932), the return appearance or second major epibole of *C. tesselatus* is noted with significant numbers in the upper fifty five feet of the Coburn (Coleville Member of Thompson, 1963). As noted by Shaw & Lespérance (1994), the genus *Cryptolithus* makes two major incursions into strata of New York. The first major occurrence begins in the uppermost Kings Falls and ends abruptly at the end of Sugar River deposition and this occurrence has been considered to contain the reference specimens for *C. tesselatus*. The genus then reappears in New York in Edenian-aged strata where forms have been referred to as *C. bellulus* or *C. lorettensis*. As noted by Shaw and Lespérance (1994) the second reappearance in New York and Ontario is dominated by morphotypes (types C & F) that are different from the dominant form in the underlying Kings Falls – Sugar River. In Pennsylvania, the genus was also noted from the Martinsburg at Swatara Gap in the southeastern Ridge and Valley (Lehmann & Pope, 1989). In the latter region forms are recognized as morph C or *C. lorettensis* (Shaw and Lespérance, 1994). Given the presence of *C. spiniferus* graptolites in the interval as noted by Lehmann and Pope (1989) as well as other data presented by Whittington (1968), these specimens appear to be latest Shermanian to earliest Edenian age. In the northwestern Ridge and Valley in the area of Bellefonte, the type area for the Salona and Coburn, Shaw & Lespérance (1994) note that *Cryptolithus tesselatus* morph A is found, but that most specimens in the Coburn show a lower number of fringe pits than forms in the Sugar River (or underlying Salona) and are more characteristic of morphotypes C and F as found in the *C. spiniferus* graptolite zone. Shaw & Lespérance (1994) expressed concern about the existing stratigraphic chronology, but commented that there appeared to be an upward change in dominance from morphotype A through C and then to F during the Shermanian to Edenian and younger. These data highly suggest that at least the upper Coburn is allied with the latest Trenton to earliest Edenian of New

York and less allied with Kirkfieldian to early Shermanian as suggested by Valek (1982) and Bellomy (1999). Hence, the Coburn is more expansive and younger than what was suggested by the inconclusive conodont data, a position that was also supported by Ross and colleagues (1982).

In addition to trilobite evidence for a younger Coburn, sowerbyellid brachiopods in Pennsylvania when compared to the type region also suggest potential for correlations. In New York Titus (1992) noted that the genus *Sowerbyella* was a fairly ubiquitous component of the Trenton and was found across a substantial range of facies. This observation is supported by statistical analysis of Patzkowsky (1995) that showed *Sowerbyella* to be a strong component of several different facies and communities ranging from the Chazyan through at least the lower half of the Mohawkian. Titus further noted that three species of *Sowerbyella* were found in the Trenton (each with at least two clinal variants). Detailed bed by bed analysis correlated over wide areas of the Trenton shelf showed each species to be relatively restricted to a specific range of strata although clinal variation between shallow and deeper water facies was recognized within a range of selected strata (Titus, 1992). Specifically, Titus recognized *Sowerbyella curdsvillensis* as the main form for the Watertown through lowest Sugar River; this species was diagnosed to have two variants (*S. c.* "*plana*" or *S. punctostriatus* of early reports that occurred in lower-energy, deeper-water environments i.e. Napanee and lowest Sugar River, and *S. c. "ornata"* that favored shallower, higher-energy environments of the Kings Falls Limestone). This form was previously recognized by Titus (1986) as *S. sericea*, and is *S. sericea* of older literature. In New York, *S. curdsvillensis* faced a "facies bottleneck" and disappeared from regions where it had previously been found during deposition of the upper Sugar River to lower Denley Limestone.

Subsequently in the overlying middle to upper Denley and into the Rust, *Sowerbyella* returned with renewed abundance and is recognized as a distinct species which Titus (1992) named *S. kayi*. This form has three variants named *S. k. "plana," S. k. "ornata,"* and *S. k. "subovalis."* Although Titus used the same variant names – there were some significant differences between *S. curdsvillensis* and its variants and *S. kayi* and its variants respectively, but again the appearance of specific forms was reflective of shallower and deeper facies. Again Titus (1992) noted that *Sowerbyella* experienced a second "facies bottleneck" in the upper Rust Formation during the *O. ruedemanni* graptolite chronozone just prior to deposition of the Steuben and Hillier Limestones which are of *C. spiniferus* age. Although this "bottleneck" was not as significant and pronounced as the first, it appeared that the shallowest form *S. kayi "subovalis"* was able to repopulate facies during the ensuing *C. spiniferus* graptolite zone and became significantly different and much greater in size from earlier forms that it had already been designated as its own species *S. subovalis* by Wilson (1932). Therefore in the Trenton of New York, Titus recognized an upward change in sowerbyellid brachiopods, both in terms of local epiboles and extinctions and an associated distinct stratigraphic distribution for these forms.

In the Ridge and Valley and in the Great Valley of Pennsylvania and Virginia, the presence of *Sowerbyella* sp. has been noted from the Salona and the Coburn Limestones and from the Martinsburg of Virginia (Thompson, 1963). As in New York, in both of the former regions, *Sowerbyella* makes at least two major appearances with evidence for substantial outages in intervening strata. Cooper (1956) noted the abundance of *Sowerbyella* in the base of the Martinsburg and again at a higher location in Virginia, and Thompson (1963) denoted the same pattern in the Ridge and Valley of Pennsylvania. Specifically, Thompson recognized an abundance of *Sowerbyella* (undetermined species) in the New Enterprise member of the Salona

Formation, and these appeared to be especially abundant in northernmost sections. The Roaring Spring Member is characteristically devoid of *Sowerbyella* although at the top of the unit and especially in the overlying Milesburg Member of the Coburn, *Sowerbyella* returns in significant numbers to produce *Sowerbyella* rudstone pavements. Thereafter, for the remainder of the Coburn, *Sowerbyella* is a significant component of the biotic constituents of the formation although its abundance decreases significantly again near the contact with the Antes Formation. This pattern is similar to and suggestive of the pattern observed in the lower to mid-Trenton of New York State.

To date, no taxonomic investigation similar to that of Titus (1992) has been completed on the sowerbyellids found in Pennsylvania, nor has any study previously described specific differences in the forms found in the Salona through Coburn to enable evaluation. Nonetheless, preliminary observations of photographs and samples of the Coburn (Milesburg Member) at Antes Gap, suggest that the forms in the Coburn are characteristically more inflated or rounded in the region of the medial fold on the pedicle and are somewhat larger in size than forms observed in the Salona. Although a detailed analysis is needed, these forms appear to be more similar to the forms described as *S. subovalis* by Wilson (1932) or potentially the form described by Titus (1992) as S. *kayi "subovalis."* If this is true, these data would suggest that the sowerbyellids in the lower Coburn are more similar to forms in the Denley and Rust Formations of New York (after the first sowerbyellid "bottle-neck") or similar to the latest forms found in the Steuben and Hillier Formations (after the second sowerbyellid "bottle-neck"). As suggested by graptolite evidence from the overlying Antes Shale, the Coburn is older at least than the medial *C. spiniferus* graptolite zone and is younger than lower *C. americanus* zone based on graptolites found in central Pennsylvania as noted earlier. Combined with the former observations, it is suggested here that substantial evidence exists for correlation of the Coburn of central Pennsylvania with the upper Shermanian of New York and as suggested by Kay (1943, 1944) the Coburn may indeed be equivalent to what was called the Cobourgian Stage in his first analyses.

Chazy Group: General Description, Contacts & Distribution

In the Ridge and Valley of central Pennsylvania, Chazy-group equivalents have been recognized on the basis of both lithologic and faunal grounds from early in the 20th century. As shown on Figure 3, Kay (1943, 1944) recognized strata including the Hatter and Loysburg Formations as Chazy Group strata. In the Cumberland Valley region, Neuman (1951) recognized the strata between the Chambersburg Limestone (Black River Group) and the underlying Beekmantown as the St. Paul Group for exposures in the vicinity of the Maryland – Pennsylvania border. As did Kay (1943, 1944), the Chazy Group equivalents of the Cumberland Valley were divided into two formations referred to as the Row Park and New Market Limestones (Neuman, 1951). The lower was subsequently divided into two units (Social Island Limestone and Browns Mills limestone members) by Swartz & Thompson (1958) (**see figure 5**). Overall the Chazy Group as exposed in Pennsylvania records the return to limestone above the significant interval of dolostone-dominated strata of the Beekmantown immediately following the Knox Unconformity. Chazy Group Limestones are dominated by a mixture of fine to coarse grained limestone facies that are typically fossiliferous especially in the upper part.

Chazy Group Lower Contact

Wagner (1963) indicated that it was “a stratigraphic custom in Pennsylvania” to place an unconformity at the base of the Loysburg (and the St. Paul Group of the Great Valley) where these units rest on the Bellefonte Dolomite and its equivalents in the Beekmantown Group. Evidence for the unconformity includes the following: 1) recognition of an irregular contact with some relief as noted especially at the base of the Row Park Formation (St. Paul Group) in the Cumberland Valley, 2) appearance of intraclastic, extraformational breccias (also in the base of the Row Park), and 3) lack of lower Chazy faunas in the Row Park and Loysburg – most are middle to upper Chazy (Neuman, 1951; Wagner, 1963). To the northwest evidence is somewhat less clear-cut.

Nonetheless, Wagner (1963) points out that the contact appears to be more transitional especially in the central and western Ridge and Valley, and that the contact is more related to lithofacies change associated with facies migration rather than unconformity. Wagner also points out that the contact is influenced by local secondary dolomitization that is often not bed-parallel. This observation counters the "erosional relief" evidence at least for this region. On this basis, it is often difficult to place the Bellefonte-Loysburg contact in the northwestern Ridge and Valley. Inasmuch as the Bellefonte is characteristically microcrystalline dolomite and the Loysburg is typically an interbedded microcrystalline dolomite and limestone unit (aka "tiger stripe"), the traditional base Chazy contact in this region appears to be a flooding style contact rather than exposure and unconformity. In contrast, the contact at the base of the Row Park and the underlying Bellefonte in the Cumberland Valley is characteristically sharp with the presence of the extraformational conglomerate and evidence of pinchout of the lowest unit of the formation (Social Island member of Swartz & Thompson, 1958). The Row Park is typically more limestone-rich (when compared to the Loysburg) with a coarser-calcarenite base and also commonly contains some quartz sand in places.

Thus as suggested by Wagner (1963), it is likely that the contact at the base of the Loysburg in the western and central Ridge and Valley is not synonymous with the basal Chazy contact. Based on lithologic evidence, Wagner suggested the Upper Bellefonte (Tea Creek of Swartz et al, 1955) and its basal sandy member (Dale Summit Member of Hunter & Parizek, 1979) may represent lower Chazy equivalent strata in central Pennsylvania, the lower of which was likely a lateral equivalent of the St. Peter Sandstone of the mid-continent region. As suggested herein, it is believed that the basal Chazy Group contact could be redefined from the top of the Tea Creek Dolomite Member of the Bellefonte to a level at the base of the Dale Summit Sandstone Member (see Figure 4). The coarser-grained, quartz-rich Dale Summit is likely the lateral equivalent of the base Row Park of the Cumberland Valley and probably coeval with the quartz-rich base of the Day Point Formation of New York. Moreover, recognition of this contact as a distinct lithologic boundary recognized on the appearance of substantially increased siliciclastic-derived sediments and sporadic to frequent extraformational conglomerates and breccias helps support the idea that this unconformity represents the Knox unconformity of the Appalachian region.

Chazy Group Upper Contact

In the Lake Champlain region of, New York the Chazy-Black River Group lithologic boundary is defined generally as the transition from variably fine-to-coarse-grained, medium to dark gray, reef and biostrome bearing limestones of the Chazy to interbedded greenish gray, sandy and argillaceous limestones and dolostones of the Pamelia Formation (Fisher, 1968). In New York, the lithologies are distinct although basal Pamelia lithologies are often included in the top of the underlying Valcour Formation making the contact somewhat gradational and difficult to place in the Lake Champlain region (Fisher, 1968). Nonetheless the Chazy-Black

River (CBR) unconformity in the southern Lake Champlain Valley and elsewhere is generally recognized where there is substantial evidence for truncation of underlying units or significant evidence for onlap by the medial Pamelia and overlying units. For instance, where the Pamelia is not present, the contact is recognized at the base of the Lowville Formation which is a mud-cracked, micritic limestone that often and displays abundant birdseye structures.

Thus over much of New York, this contact is unconformable with significant evidence for truncation of underlying units including lowermost Pamelia (with truncation of Chazy over large regions). In the northern Lake Champlain region, where the gap is significantly less, the Pamelia shows a lower and upper unit separated by a recognizable unconformity. The lower unit of the Pamelia is argillaceous and dolomitic and is distinguished from the underlying Valcour on the basis that the Valcour is fossiliferous, and only occasionally dolomitized in some cross-stratified zones (Oxley & Kay, 1959). In this same region, the upper unit of the Pamelia contains a somewhat intraclastic base with fine to coarse calcisiltites that grade upward to medium-to-massive, cross-bedded sandy dolomitic limestones and dolostones. This upper unit is the most common facies of the Pamelia Formation, especially where it oversteps and truncates underlying units.

In central Pennsylvania, the top contact of the Chazy has been variably identified since the term was first applied to Pennsylvania strata. Kay (1944) was the first to formalize a specific connection to the Chazy of the type region, but see D'Invilliers use of the term (f**igure 3**). Prior to that time, most stratigraphic comparisons (Butts, 1918; 1931; and Field, 1919) were based on the suspected equivalency of this interval to the Stone's River Group of Tennessee. Local units were recognized and a number of internal divisions were distinguished within the "Stone's River," but it was argued by Butts (1931) that the Stone's River in Pennsylvania could generally be divided into two major units – an upper or Black River Limestone and a lower or Carlim Limestone which was considered Chazyan – although the stratigraphic term was not used by him locally. Kay (1944) recognized this same division, but favored application of the New York-based nomenclature for both units and therefore installed Chazy Group nomenclature for the interval below the Black River.

Kay formalized the nomenclature and used Hatter for the upper formation, and used Field's (1919) Loysburg Formation for the lower. As did Butts (1931), Kay (1944) recognized the contact between the Loysburg Formation and the overlying Hatter (Butt's Lemont Limestone) as disconformable. He also recognized the contact at the top of the Hatter as unconformable as well. Given the pronounced Pamelia unconformity in New York, Kay felt there was evidence in the base of the Black River of Pennsylvania (at the base of the Snyder Member) for an unconformity as in New York. Kay identified the upward increase in argillaceous content of the Hatter Formation as representative of the transition from the Valcour into the Pamelia and considered the contact at the top of the Hatter as roughly the equivalent of the CBR unconformity. Moreover, the occurrence of "dolomitic beds" and mud-cracked intervals in the overlying Snyder suggested more restricted facies typical of the Pamelia of New York which helped support Kay's model.

Nonetheless, subsequent work by Swartz, (1955) favored the Loysburg-Hatter boundary as the most significant unconformity represented in Pennsylvania. Thus Swartz considered the CBR unconformity of New York as synonymous with the Loysburg-Hatter contact, and included the argillaceous Hatter Limestone within the overlying Black River despite the faunal and lithologic challenges presented by Kay. Rones (1955, 1969) and subsequent workers have continued to use the model of Swartz (1955) where the Hatter Limestone is considered lithologically more similar to the overlying Black River units than to underlying Loysburg (see **figure 3).** This is also reflected in the recent work of Laughrey and colleagues (2003, 2004), even though other biostratigraphic work on brachiopods and ostracods (Cooper, 1956; Swain, 1957), as well as onlap-offlap curves from siliciclastic residues (Rones, 1969), suggested the Hatter was at least partially equivalent to the Chazy and that the Hatter-Snyder contact was indeed unconformable.

Herein these data are considered strong evidence in support of a somewhat modified scenario following more closely the original designation by Kay (1944) for the rough position of the Chazy-Black River lithostratigraphic boundary (see Figure 6). Moreover, the contact at the base of the Hatter although also an unconformity is thought to be equivalent to another unconformity recognized internally within the Chazy. It is likely that the CBR unconformity, positioned at the top of the lower Pamelia discontinuity of the northern Lake Champlain Valley, actually correlates with the Hatter-Snyder unconformity of current usage. In Pennsylvania, the actual CBR lithofacies boundary might be drawn at a somewhat lower level which probably corresponds to a position below the top of the Hostler Member and within the upper Grazier Member. Nonetheless, in order to maintain original lithofacies designations for each of the formations, the Chazy Group in Pennsylvania should be considered inclusive of strata through the top the Hostler Member of the Hatter Limestone with the base of the Black River Group initiating at the base of the Snyder Formation rather than at the base of the entire Hatter Formation.

Distribution:

As defined, the Chazy Group (Loysburg and the majority of the Hatter formations) can be recognized throughout most of the Ridge and Valley and in the Cumberland Valley. Swartz (1948) recognized significant thicknesses of the Chazy interval within a broad northeast-southwest trending basin in the subsurface of Pennsylvania which he referred to as the Appalachian Trough. This basin was positioned from just west of the Alleghany Front eastward to the southeastern edge of the Great Valley. In westernmost Pennsylvania, Swartz (1948) inferred that Chazy Group strata were either very thin or absent. This area of thinning was also noted by Ryder and colleagues (2001) and is coincident with the western margin of the Pre-Cambrian Rome Trough rift basin – the northward extension of which is known as the Olin Trough (**figure 8**). In a transect from West Virginia into Ohio near the Pennsylvania border,

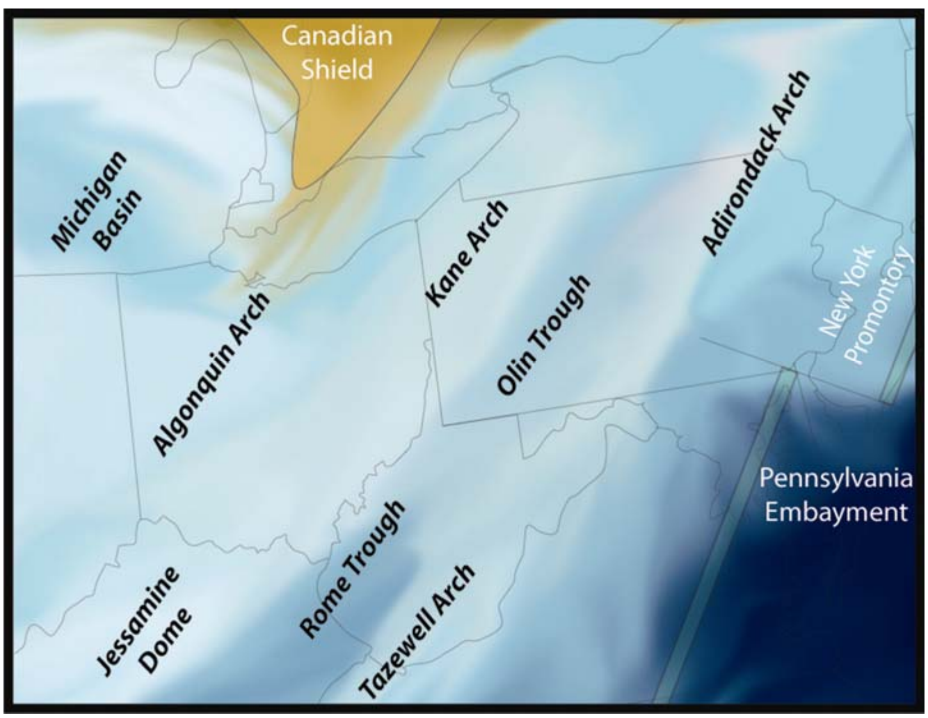


Figure 8: Paleogeographic reconstruction for the east-central Great American Carbonate Bank in the region of Pennsylvania and the central Appalachians leading into the Taconic Orogeny.

data from Ryder (1991) demonstrates a considerably thinned St. Paul Group (Chazy equivalents of the Great Valley). West of the Olin trough and the Kane Arch, these authors correlate the St.

Paul with the Wells Creek interval which is recognized as their "unnamed argillaceous limestone member" of the St. Paul.

The Chazy Group is thickest in south-central Pennsylvania near the Maryland border in the Cumberland Valley (Neuman, 1951) along the southeastern margin of what Kay (1944) referred to as the Adirondack Axis, but thins to the west across the arch into the western portions of the Ridge and Valley. Based on stratigraphic relationships with the Jacksonburg Limestone and underlying Beekmantown, in the regions to the east of the Cumberland Valley (i.e., in the Lebanon Valley and certainly in the Lehigh Valley) Chazy strata are absent by unconformity beneath the Jacksonburg Limestone which is of early Trenton age (Hobson, 1963). Read (1989) attributed this eastward thinning to uplifted areas along the New York Promontory or what he referred to as the New Jersey Arch. The thickest Chazy carbonates are found in the vicinity of the Pennsylvania Embayment and south of the Tyrone-Mt. Union Lineament that runs roughly perpendicular to the trend of the basins and arches as previously recognized. South of Pennsylvania, Chazy strata thin in the southern Shenandoah Valley and in eastern West Virginia before becoming thicker again in southwestern Virginia in the position of Reed's (1989) Virginia Arch. In the subsurface to the north of the Ridge and Valley, Wagner (1966) shows the Chazy equivalents to become truncated under Black River strata just north of the New York border – a pattern also shown by Fisher (1977).

Thus as pointed out by Wagner (1966) and Reed (1989), the central to southern Pennsylvania area of the Appalachian Trough (of Swartz, 1948) was a region of significant carbonate deposition both along Kay's Adirondack Axis and to the east of the Olin Trough. Deposition in this region was strongly related to ancestral basement structures associated with the New York Promontory / New Jersey Arch, the Pennsylvania Embayment / Tyrone – Mt.

Union Lineament, and the Virginia Promontory / Tazewell Arch. Distribution of Chazy Group deposits was thus likely influenced by both tectonic activation of the GACB margin during the Tinmouth Phase of the Taconic Orogeny (Rodgers, 1971) and sea-level change that flooded the passive margin during the Tippecanoe megasequence transgression.

Chazy Group Formations

Dale Summit Member of the Bellefonte Dolomite

The Dale Summit Sandstone was formally recognized by Swartz and colleagues (1955), who indicate that it was first mapped by Butts (1918) in the Nittany Valley between Dale Summit and Bellefonte in the western Ridge and Valley. The unit is relatively thin (ranges up to ~7 meters) and discontinuous but appears to have been deposited in depressions on the top of the underlying Coffee Run Dolostone Member of the Bellefonte. The Dale Summit is described as a fossiliferous, quartz-rich sandstone with well-to-perfectly rounded quartz sands (Swartz et al., 1955). In some outcrops the unit develops into a conglomerate with clasts of limestone, dolostone, and chert – evidently derived from the underlying Coffee Run Member of the Bellefonte. Wagner (1966) notes the presence of the member in the subsurface of the Ridge and Valley in southern Pennsylvania and into northern West Virginia where the unit becomes somewhat thicker. In the subsurface of Bedford County, Pennsylvania the well-rounded, quartz sandstone is shown to be about 3 meters thick and is also associated with some thin greenish-grey shale beds. Fossils in the Dale Summit include unidentifiable gastropods and fragments of trilobites and were thought by Kay (1944) to possibly represent evidence for the first faunas of the post-Knox Chazy Group.

Tea Creek Member of the Bellefonte Dolostone

The Tea Creek Member of the Bellefonte Dolostone is an unfossiliferous, microcrystalline to very fine-textured dolostone and is recognized from the northern end of the Ridge and Valley southward in outcrop and core to the Bedford County region where it begins to interfinger with micritic limestones (Wagner, 1966). In outcrop the Tea Creek is medium-gray in color on fresh surfaces with reddish variegated coloration that helps identify some minor bioturbation (Swartz et al., 1955). The unit weathers to a whitish yellow gray color and tends to be highly etched in outcrop although it is commonly covered in valley exposures. The Tea Creek is contrasted with the underlying Coffee Run Member of the Bellefonte Dolostone on the basis that the Coffee Run is often much more coarsely dolomitic and significantly browner in color on fresh surfaces. In weathered sections the beds display a more characteristic buff brown. Overall in the Nittany Valley region, the Tea Creek is about 60 meters thick, but thins to the northwest (Swartz et al., 1955). In the subsurface of northwestern Pennsylvania in the region of the Kane Arch, the Tea Creek becomes very thin (to less than 15 meters), and is associated with an interval of red-to-green shales that are referred to as the Shadow Lake Formation based on correlations into Ontario (Wagner, 1966). Wagner believed that the Shadow Lake was a lateral equivalent (at least in part) of the Tea Creek Member, but the Shadow Lake may also be in part younger than the Tea Creek.

Loysburg Formation:

The Loysburg Formation was first defined by Field (1919) for exposures of dark, siliciclastic-influenced, dolomitic limestones in the southwestern Ridge and Valley in Bedford County. Faill and colleagues (1989) indicate that the Loysburg extends throughout Ridge and

Valley in outcrop and can be traced in the subsurface to just past the New York border before it is truncated (as per Swartz, 1948). As originally defined, the Loysburg was a 35 meter thick interval between the Beekmantown (Tea Creek member of the Bellefonte) and the first intraformational conglomerate zone of the Upper Stones River or Carlim Formation (now Hatter Formation). Field identified the unit as distinct not only on the basis of its lithology, but also on the absence of any reef-building organisms that were found in the overlying units. Kay (1944) subdivided the unit into two members – the lowest was referred to as the "Tiger-Striped" member, and the upper as the Clover Member. Swartz and colleagues (1955) subsequently substituted the name Milroy Member for the "Tiger-Striped" member.

Milroy Member

The lower Loysburg Milroy Member is a distinctive unit that might be called a ribbon carbonate by some workers, but is composed of 2.5 to 5 centimeters-thick, dark, coarse-grained, wavy-bedded, dolomitic beds interbedded with finer-grained limestone ribbons of about the same thickness. The "striping" weathers in relief on exposed surfaces. North of the type region, the Milroy thickens substantially so that in the vicinity of Bellefonte, over 90 meters of Milroy is reported to be present (Wagner, 1955). Chafetz (1969) specifically defined the lower contact at the base of the first limestone above the highest dolostone of the Tea Creek and placed the upper contact at the base of lowest bed showing abundant and diverse fauna. Rones (1969) felt that the base of the Loysburg as defined could be demonstrated as unconformable – a position held also by Kay (1944). Internally, the ribbon limestone beds are often laminated and occasionally demonstrate algal crinkle-lamination. Swain (1957) described the basal unit in exposures at Union Furnace and indicated the presence of a few minor breccias and algal (*Solenopora*?) bioclastic beds. Swain also described the lower part of the Loysburg as containing minor ostracod-brachiopod-bearing beds and occasional shaly-textured dolomitic limestones. These shaly zones contained a few ooids in a thin zone (~30 cm) and noted where some ooids had apparently fallen into vertical tubes – likely burrows. He also noted the occurrence of a zone stromatolites (algal heads) with leperditian ostracodes and bumastid trilobites in the middle of the Milroy which were followed by another bioclastic "algal" breccia. The top of the unit returns to more typical "ribbon" dolostones and limestones beds.

Clover Member

The upper member of the Loysburg, the Clover, is a sublithographic medium-to-dark gray ostracod-bearing limestone with some medium dark-gray silty argillaceous beds. Kay (1944) described the Clover as ranging in thickness from 12 to 24 meters with about 20 meters at Union Furnace. Swain (1957) recognized a slightly greater thickness than Kay in the Union Furnace cuts, and included about 30 meters in the Clover member. The difference is difficult to reconcile. However, it looks as though Swain defined the base of his Clover to the level of the first ostracod beds. In contrast to the underlying Milroy member, the unit is commonly medium to massive-bedded and lacks the uniform interbedded dolostone ribbons and often weathers white and contains a couple of intraformational conglomerates. Kay recognized the more dolomitic intervals with conglomerate/breccia zones as they are typically darker in color and silty and occasionally show some evidence for channeling. In general, the more massive nature of this unit is due to extensive bioturbation and amalgamation of bed boundaries.

Wagner (1966) indicated that it becomes difficult to separate the Clover from the underlying Milroy in the subsurface, but in northwestern Pennsylvania, the units are separated based on recognition of the silty-argillaceous zone recognized in outcrop. This silty-argillaceous zone appears to be the same interval as Kay's dolomitic, conglomerate interval and corresponds to Swain's argillaceous ooid bed. The more massive sub-lithographic beds become more prevalent above this level and are occasionally interbedded with thin argillaceous beds, thus assigning this unit to the base of the Clover seems to be a reasonable solution. Nonetheless, Faill and colleagues (1989) report the thickness of the Clover along PA Rte 453 at Union Furnace to be only about 6 meters thick. It is clear that these authors have included significantly less than was originally intended. Herein we follow the amended usage of Kay and Swain's concept for the Clover. Based on ostracod zonation established by Swain (1957), the interval to the top of the Clover Limestone has been included within the *Aparchites kaufmanensis* zone, thus correlating the Loysburg Formation with the lower to middle Chazy of New York State (Day Point to Crown Point Formations).

Hatter Formation:

Despite recognition by Kay (1944) that the Hatter Formation in central Pennsylvania represented the uppermost Chazy Group, subsequent authors included the fossiliferous Hatter with the overlying Black River Group equivalents. As discussed previously, this writer favors the original classification of Kay based on numerous lithostratigraphic and biostratigraphic arguments including those suggested by Swain (1957). Therefore, the Hatter Formation is included within the Chazy Group although the actual contact might be within the Hatter (**see figures 6 & 7**). The Hatter Formation was named for Hatter Creek in the vicinity of Roaring Spring, Pennsylvania. However Kay's type-section is at Union Furnace. Kay (1944) defined the unit to contain three members from base to summit to include: the Eyer, Grazier, and Hostler Members. Overall, the Hatter Formation is significantly enriched in siliciclastic components including siliciclastic clays and detrital quartz grains that can make up as much as 36% of the insoluble rock matrix (Kay, 1944; Berkheiser & Lollis, 1986) especially in the upper members of the formation.

In the western Ridge and Valley, the overall thickness of the Hatter ranges between 21 to 30 meters (Kay, 1944; Wagner, 1966) from north to south respectively. The Hatter also thins to less than 18 meters toward the east into Path Valley before thickening again and transitioning into the Shippensburg Limestone of the Cumberland Valley. The Hatter is generally very easy to recognize on the basis of its sharp basal contact with the Clover Member of the Loysburg. In this case the coarse-grained basal member of the Hatter (where present) stands in sharp contrast to the sub-lithographic or calcilutites of the Clover. The top of the formation is easy to recognize using the lowest occurrence of ooid grainstones from the base of the overlying Snyder Formation. In the subsurface, the Hatter is recognized in West Virginia and is delimited between the base of the overlying Snyder and the top of the Clover – although internal differentiation is not deemed possible as per Wagner (1966).

Eyer Member

The Eyer was described as the lower granular limestone member of the Hatter Formation (Kay, 1943, 1944). The Eyer is relatively thin to absent- especially in northern portions of the Ridge and Valley (i.e. northern Nittany Valley), but at Union Furnace the thickness of the grainstone is about two meters where it consists of medium-dark gray calcarenite that transitions upward into another meter and a half of thin interbedded fine-grained dolomitic limestones (calcilutites) with thin shale stringers. Faill and colleagues (1989) report 3.4+/-0.4 m for the unit at Union Furnace, which is in agreement with measurements reported here (**see figure 7**). The unit is more fossiliferous than the underlying unit and contains occasional large colonies of the tabulate coral *Tetradium fibratum*. As shown by Kay, the Eyer sharply overlies the Loysburg with the contact placed at the base of the first fossiliferous calcarenite – a contact which he considers to be disconformable. The Eyer member has a diverse fauna relative to the underlying Loysburg strata, nonetheless the assemblage is still a restricted fauna dominated by thin branching stictoporate bryozoans, leperditian ostracods, a few species of brachiopods, including the orthid *Pionodema* that is occasionally abundant, as well fragments of *Bathyurellus* (Kay, 1944). In addition to fossil beds, the Eyer also contains the lowest K-bentonite recognized in the Union Furnace outcrop – informally recognized as B-0 (Berkheiser and Lollis, 1986).

Grazier Member

The Grazier Member is a medium-gray, fairly massive bioturbated wackestone to mudstone lithology with thin argillaceous partings, which Rones (1969) referred to as a gray, medium- to-thick-bedded calcisiltite and calcilutite interval, although the calcilutite lithology can be dominant (Wagner, 1966). Kay (1944) indicated that the unit is typically about 10 to 11 meters thick, but is somewhat thinner in the Nittany Valley where the Grazier is in contact with the underlying Clover. To the southeast, the Grazier thins down to less than 3 meters in the Path Valley. Faill and colleagues report a thickness for the Grazier Member at 14.6+/-2 m at Union Furnace. The basal bed of the bioturbated wackestones contains a rippled top with vertical chert-filled burrows and appears to be a good regional marker. The massive lower interval (~10 meters) grades sharply upward across a dark-stained, mineralized flooding-style surface (possible Chazy-Black River lithofacies boundary?) into an interval of wavy-to-planar interbedded calcilutites and calcisiltites that are significantly less bioturbated, but are substantially darker. These beds grade upward into more massive, bioturbated wackestones that are fossiliferous with large cephalopods (*Gonioceras* sp.), and small colonies of bryozoans. The unit also contains fragments of what appear to be receptaculitid algae.

Kay (1944) viewed the entire Hatter Formation as predominantly a transgressive interval with minor evidence for regression at some horizons. Thus he believed the base of the Grazier at the contact with the underlying Eyer to represent a disconformity especially where the underlying Eyer was truncated or “overlapped.” Interestingly, Berkheiser & Lollis (1986) indicated a significant upward increase in insoluble siliciclastic detritus in this interval of the Hatter reaching nearly twenty percent insoluble components (mostly clastics) by the upper contact of the Grazier just above the mineralized surface. This rapid upward surge from values less than about five percent below the mineralized surface suggests a major influx of siliciclastic sediment during a period of flooding.

Hostler Member

The upper member of the Hatter Formation is a substantially more siliceous limestone and is referred to as the Hostler Member. Overall, it is coarser-grained than the Grazier and exhibits limestone facies which are very similar to portions of the Crown Point and Valcour Formations of the type region except with substantially more siliciclastic material. As suggested by Faill and colleagues (1989), the unit is internally differentiated into sub-units. As shown on **figure 7** the lower unit consists of wavy-bedded wackestones and calcisiltites that grade upward into slightly dolomitic, laminated calcisiltites. These are sharply overlain by an interval of fossiliferous coarse-grained, rippled calcarenites, packstones, and bioturbated wackestones. This middle package in turn fines back to dolomitic lutites with occasional *Tetradium* and relatively high siliciclastic material that is typical of the Black River. The green and planar-bedded dolomitic siltstone at the top of the lower interval (mining unit 6 of Berkheiser & Lollis) contains up to 36.2 percent insoluble residues some of which appear to be disseminated cherts and quartz silt (Berkheiser and Lollis, 1986). This interval may represent the Lower Pamelia equivalent. Above this horizon insoluble residues show lowered values to a low for the interval (mining unit 9) at 7.9 percent at the top where it is sharply overlain by the extraformational conglomerates/breccias and ooid bearing intervals at the base of the Snyder Formation (Berkheiser & Lollis, 1986).

In the subsurface, Wagner (1966) discriminates an upper and lower interval respectively and documents the upper four meters to be representative of the dolomitic calcilutite facies (somewhat like the underlying Loysburg), but he doesn't further differentiate the unit. In outcrop Faill and colleagues (1989) report the member overall to be about 12+/-2 m thick in the vicinity of Union Furnace. Herein we report that the interval is just over 11.5 meters at Union Furnace and are thus in agreement.

Faunally, the Hostler Member shows the greatest diversity of all the Hatter Formation members. Kay (1944) reports at least 28 different taxa (compared to 18 of the underlying units) with a much greater diversity in types of cephalopods, brachiopods, trilobites, bryozoans, and corals. Kay reported that the faunas of this unit were mixed in that the assemblage contained some genera and species found in the Black River while others were from the Chazy. He thus believed both on lithologic and biostratigraphic basis, that the Hatter Limestone was transitional to, but older than the majority of the Black River Group and younger than the majority of the Chazy Group. It is clearly the most open marine (although still protected) facies represented thus far in the Upper Ordovician of central Pennsylvania and clearly similar to facies of the Chazy. Kay (1944) also suggested the Hatter to be equivalent to the Ottosee of the southern

Appalachians based on faunal content and presumably based on increased siliciclastic composition. Moreover, as recognized by Rones (1969) the upper contact of the Hatter Formation with the Snyder Formation is sharp and disconformable with clasts of the Hatter and even potentially pieces of Loysburg included in the Snyder. This contact is likely equivalent to the CBR unconformity discussed earlier.

Black River Group: General Description, Contacts & Distribution

In the Ridge and Valley of central Pennsylvania, Black River Group equivalents have been formally recognized beginning early in the 20th century based on both lithologic and faunal grounds. As shown on **figure 3**, Collie (1903) used the term Black River Limestone in Pennsylvania. As was originally defined, the Black River Limestone was only the top portion of what has today become an extended Black River Group synonymous with the older term "Birdseye Limestone." In the early studies, the Black River was characterized as a fossiliferous, dark grey to black, massive-bedded, fine-to-medium grained limestone with occasional dark grey to black cherts. The Black River was also recognized as the massive limestone unit that occurred just below the interbedded limestones and shales of the Trenton Group. As in New York subsequent revision of the Black River, to include the fine-grained fenestral (birdseye) micrites and associated peritidal deposits of the Lowville and Pamelia Formations (as per Cushing et al., 1910), resulted in the expansion of the unit to include lower strata.

In Pennsylvania Butts (1918, 1931) and other workers constructed a number of local formation or member level names, but continued to include them within the concept of the expanded Black River Group. Butts (1918) defined an upper and lower interval – recognizing the lower as Lowville Formation equivalents and the upper as the Rodman Formation. Butts'

Rodman represented the interval considered to be Black River by early workers including Collie (1903). Today the Rodman is used in a restricted sense as a member of the Nealmont Formation and includes only a portion of its original thickness. Following work by Field (1919) and Kay (1943, 1944) further lithologic and biostratigraphic developments resulted in modification of earlier interpretations. Based on correlations to the Ordovician of New York, Kay defined the Black River Group to include the interval from the base of the Snyder Formation upward to the base of the Nealmont Formation (**see figure 7**). Since he believed the Nealmont to be Rocklandian and Kirkfieldian in age, Kay (1944) removed the Nealmont from the Black River as was the tradition, leaving the Black River to include only two formations - the Benner (essentially Linden Hall of current usage) and Curtin Limestones. Thus he excluded the portion of rocks considered by early workers to be representative of type Black River Limestone.

Swartz (1955) followed Kay's usage and stratigraphic nomenclature, while Rones (1955, 1969) considered Kay's Curtin Formation to be lateral facies equivalents of the upper Benner Formation. Thus Rones dropped the Curtin Formation, redefined the Benner to include the members of Kay's Curtin Formation, applied the new name Linden Hall to the interval, and promoted the basal Benner unit (Kay's Snyder Member) to formation status. As shown in his posthumous publication, Rones' (1969) manuscript defined the Black River to include the Linden Hall, Snyder and Hatter Formations thus departing from the usage of Kay. Cognizant of the discrepancy and recognizing the ongoing debate regarding the Black River Group, Wagner (1966) referred to the interval as the "middle carbonate unit" in his investigation of the subsurface of Pennsylvania. Nonetheless in order to accommodate both Kay's and Rones' ideas, he correlated the Nealmont with the Kirkfield of New York and Ontario, the Benner (Linden Hall) with the Chaumont, Rockland, and Bobcaygeon Formations of New York and Ontario; the

Snyder with the Upper Lowville and the Hatter with the Lower Lowville and Pamelia of New York. This usage excluded those strata that were originally classified as representative of both Lowville lithologies and faunas by Butts (1918, 1931), and clearly represented the extreme diachronistic Waltherian model for deposition of both facies and faunas that challenged earlier "layer-cake" models.

Thus without any major new re-investigation of these intervals, since the late 1950s, classification of the Black River Group in Pennsylvania follows the expanded models proposed by Rones (1955, 1969; see for example Laughrey et al., 2003, **as shown in figure 3 above**). However, based on recent lithostratigraphic, and biostratigraphic assessments referenced herein, this work favors a modified usage for the Black River Group. That is the Black River Group lithologically should include the Snyder and Linden Hall Formations, and exclude the Hatter Formation as suggested by Kay (1944). Moreover, it is recognized herein, that the Nealmont Formation, which was originally classified as the Black River Limestone is likely the equivalent of the Watertown and Selby Formations of New York and Ontario. Given that the Black River-Trenton (BRT) contact is drawn at the base of the Selby Formation (**see figure 6**), the BRT boundary likely sits at the base of the Rodman member of the Nealmont. Nonetheless in order to retain lithostratigraphic concepts for the Nealmont Formation at this time, the Nealmont is included in its entirety as Trenton following earlier precedents – however recognizing the age of the Nealmont as Turinian to early Rocklandian and not Kirkfieldian as proposed by Kay (1944).

Black River Group Contacts

As defined previously, the contact at the base of the Black River Group in Pennsylvania is established at the top of the Hatter Formation and further discussion is redundant and unnecessary here. In turn, the upper contact of the Black River is currently established following the work of Kay (1944) and by most subsequent authors at the base of the Nealmont Formation. Both Rones (1955, 1969), and Kay (1944), recognized a contact at the base of the Nealmont that truncated the upper Linden Hall Formation to the southeast and south of the Nittany Valley region. Although this unconformity was not identified in the subsurface, Wagner (1966) nonetheless recognized the transition from massive bioturbated microcrystalline to very fine-textured limestones (micrites to wackestones) of the Linden Hall (top of his "Middle Carbonate Unit" to more argillaceous, more fossiliferous beds of his "Upper Carbonate Unit" which included the Nealmont, Salona and Coburn Formations. Toward northwestern Pennsylvania along the Kane Arch, the transition beds interfinger with calcarenites similar to the Coboconk of Ontario. Wagner (1966) partially correlated these beds with the Bobcaygeon and recognized that they lay between the Gull River and the Verulam of Ontario.

Kay (1944) and Rones (1969) presented the details for recognition of the basal Nealmont unconformity. Nonetheless as pointed out by Rones (1969) there are several points of disagreement as they recognize the contact differently (**figure 9**). As suggested by Kay (1944), the base of the Nealmont was placed at the base of the Oak Hall Member of the Nealmont (the lowest of his three members of the Nealmont). In the type section of the Nealmont at Union Furnace, Kay recognized a one meter thick coarse-grained limestone (calcarenite) sitting above the Stover-Valley View members of his Benner Formation (lower members of the Linden Hall Formation). He considered the base of the grainstone to be the contact between the Trenton and Black River groups. Elsewhere to the north in the Bellefonte to Pleasant Gap area of the Nittany Valley, the base of the Nealmont was positioned at the base of the Centre Hall (middle Nealmont Member) where it rested unconformably on the Valentine and Valley View Members of the

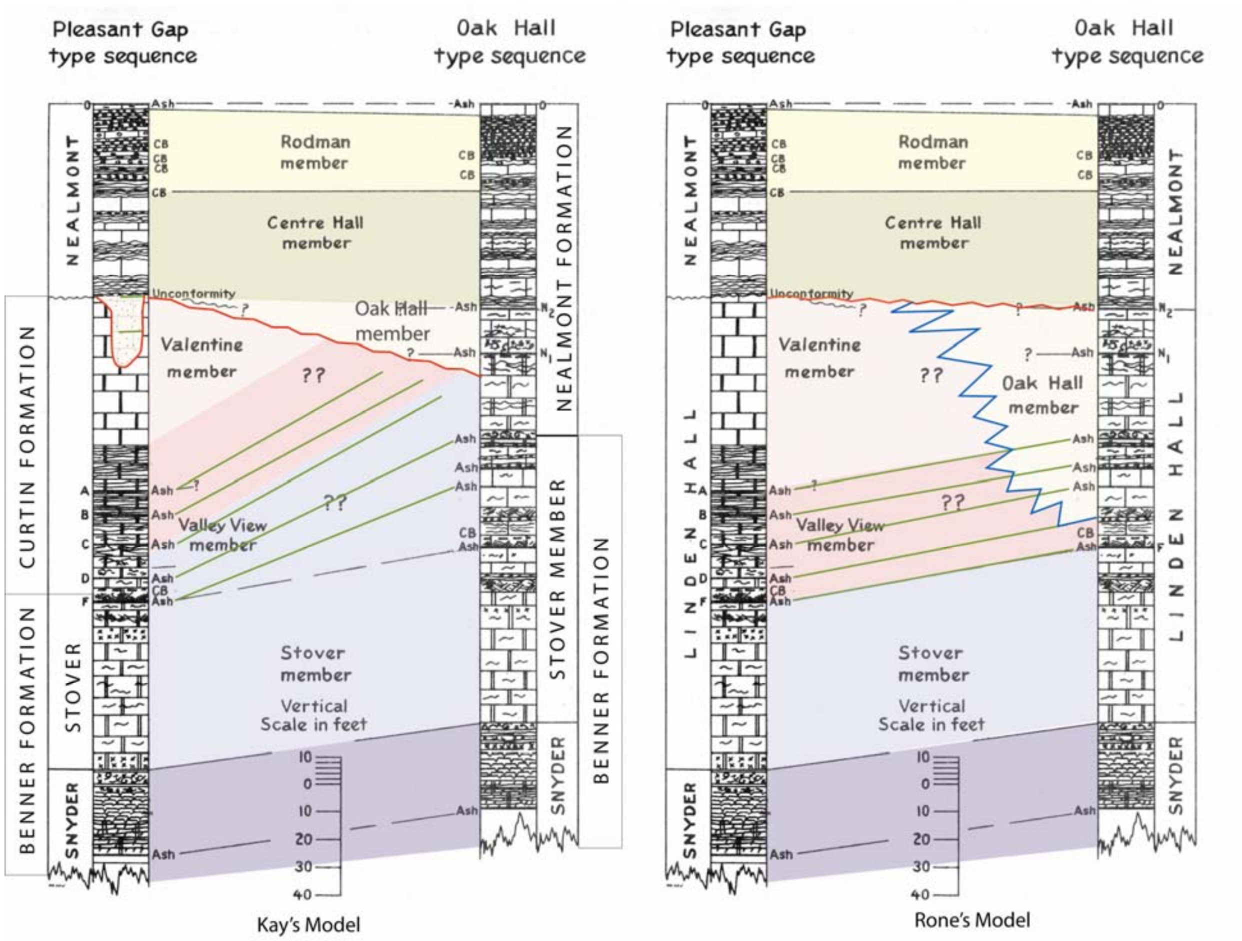


Figure 9: Correlated stratigraphic columns between Pleasant Gap and Oak Hall (a distance of 11 km) modified after Rones (1969). Columns on the left depicted to represent the stratigraphic model proposed by Kay (1944). This model identified the Black River - Trenton unconformity(in red) at the base of the Oak Hall Member of the Nealmont. Also identified is a channel feature as described by Kay from the Colerain Quarry north of Bellefonte. The presence of a channel with bentonite material has intermittently been exposed during blasting in the active quarry at Pleasant Gap but is discontinuous as described by quarry workers. On the right is the model proposed by Rones (1969) which considers the Oak Hall as a lateral equivalent of the upper Valley View and Valentine, with the principal Black River – Trenton unconformity located at the base of the Centre Hall member of the Nealmont and at the top of an expanded Linden Hall.

Linden Hall Formation (Kay's Benner). In these locations the basal coarse-grained Oak Hall Limestone was not present. However, eleven kilometers southwest of Pleasant Gap in the vicinity of Oak Hall, the coarse-textured limestone reappeared, but was of variable thickness ranging from absent to two or more meters in exposed sections south and southeast of Nittany Mountain. North of Bellefonte, Kay recognized an area of substantial channeling (karst?) into the Valentine Limestone of the Linden Hall. The overlying Nealmont filled in the channels and was characteristically coarse-grained and similar to the Oak Hall lithologies, but was capped by the Centre Hall. In the quarry at Coleville just north of Bellefonte, Kay (1944) described the succession of rocks in the channel below the Centre Hall. In this scenario Kay described the Valentine to contain an abnormal occurrence (~3.6 meters) of fossiliferous intraformational conglomerate at the top of the characteristic massive pure chemical lime that characterizes the unit. In a wedge shaped interval above this, he noted about 50 centimeters of a tough limestone overlain by 2.29 meters of fossiliferous calcarenite with *Columnaria halli* coral colonies and other coarse fossil debris. This in turn was overlain by 2.12 meters of light grey to yellow weathering ribbon limestones with argillaceous partings that contained “flow structures” (possibly seismites?), and desiccation cracks. Finally the entire succession was capped with 5.5 meters of typical Centre Hall lithologies that graded up into the Rodman. Kay recognized this interval as somewhat distinct for the Oak Hall but considered it to be representative of the upper Oak Hall at Oak Hall to the south. As this quarry subsequently closed and is now significantly overgrown and inaccessible, it is impossible to investigate the section described further. Nonetheless, discussions with quarry operators at the Glenn O. Hawbaker Quarry on the south side of the valley at Pleasant Gap indicated the presence of an intermittently developed coarse unit with thick beds of K-bentonite at the top of the Valentine. During quarry operations, operators periodically remove the channel-filled lenses of K-bentonite when they are exposed suggesting similarities to what was described by Kay.

In contrast, Rones (1969) favored a model in which the Valley View and Valentine Members were lateral facies and time equivalent to the Oak Hall Formation as shown in **figures 9 and 10**. Rones recognized a small unconformity above the Oak Hall at the base of the Centre Hall that was responsible for beveling both the Oak Hall and the Valentine. At Pleasant Gap, the

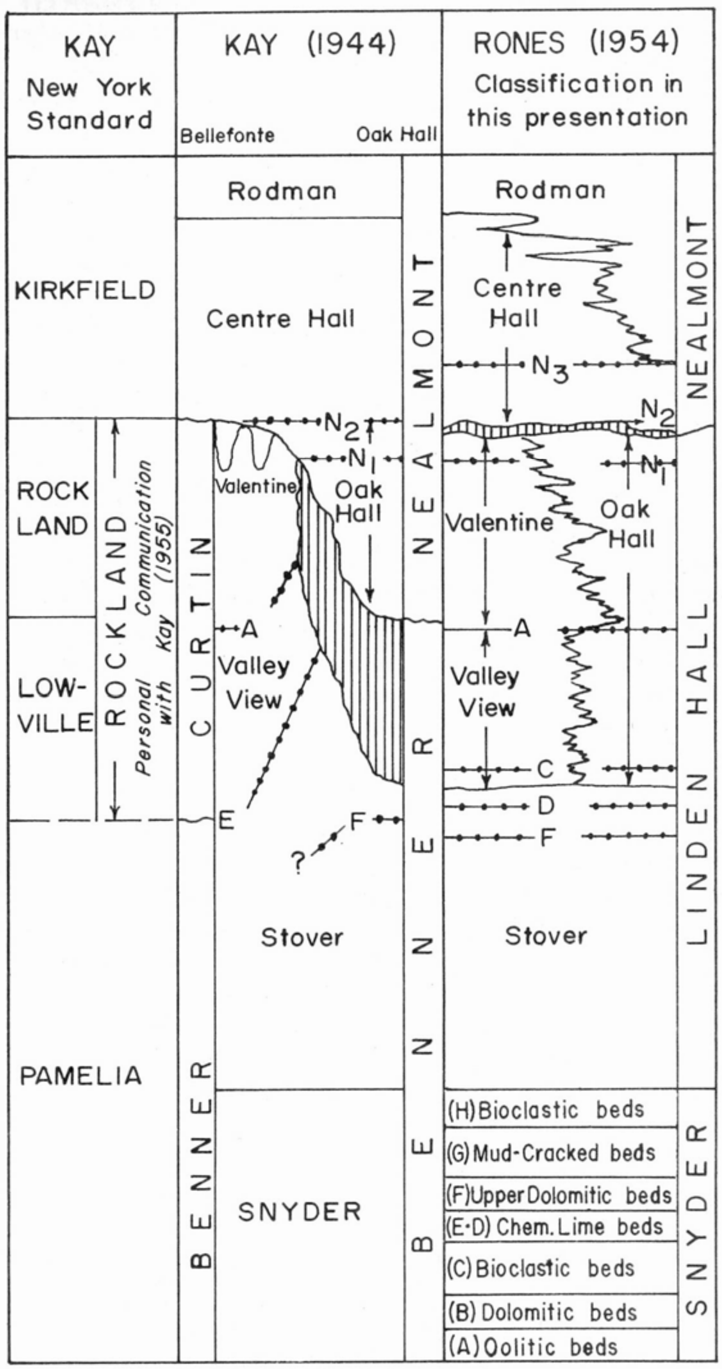


Figure 10: Stratigraphic models for the Black-River-Trenton Unconformity in Pennsylvania as presented by Rones (1969). Showing the relative extent and position of the Nealmont Unconformity as described by Kay (1944) versus that favored by Rones (1969)

N3 K-bentonite in the Centre Hall is clearly developed but the section lacked the lower N2 and N1 K-bentonites and the Oak Hall Formation which Rones suggested were removed by unconformity at this location, although they appear to be intermittently exposed today during quarrying operations. Further north near Salona (~32 km north), Rones recognized the occurrence of a K-bentonite in beds of light-grey fine grained calcilutite facies (much like the Valentine further south) which were inter-fingered with coarse-grained calcarenites similar to the Oak Hall. The shaly Centre Hall was located less than three meters above this level which suggested this was either the N1 or N2 K-bentonite. Further east, the K-bentonite swarm in the Valley View Formation (K-b's A-E) is not recognized and only a single K-bentonite (Bentonite F of Rosenkrans, 1934) is recognized in the Stover member of the Linden Hall in the southeastern valleys.

Nonetheless several K-bentonites are recognized in the Oak Hall to Centre Hall interval in the Kishacoquillas Valley both at Reedsville (N1, N2, and N3 are recognized) as well as at Naginey where the N1 and N3 are recognized (Rones, 1969) with only limited evidence for discontinuity between the Centre Hall and the underlying Oak Hall. Thus the majority of Rones' evidence came from 1) the recognition of the (N1?) K-bentonite in interbedded facies similar to the Valentine, 2) recognition of facies inter-fingering in the northeastern-most Nittany Valley, and 3) the view that the Oak Hall was discontinuous and truncated from above beneath the Centre Hall. In Rones' view the most significant unconformity was thus at the base of the Centre Hall although he suggested in a couple of locations that the Oak Hall could potentially show some evidence of basal unconformity as well.

Collectively this information suggested to Kay that the Nealmont represented a transgressive deposit over a regional unconformity with the Oak Hall deposited only in channels and low topographic areas developed on the exposed and eroded surface of the Linden Hall Formation. While for Rones, the recognition of some lateral facies change in some units suggested that the Valley View and Valentine were lateral equivalents of the Oak Hall (which appears to be an expanded unit compared to what was recognized by Kay) with all three sitting below the unconformity at the base of the Centre Hall.

Given the occurrence of *Maclurites logani* and other upper Black River taxa, together with the lithology of this unit and the chert-rich intervals in the Oak Hall and Centre Hall, the pattern is very similar to the succession found at the top of the Lowville Formation of New York leading into the Watertown Limestone. This interval in New York is referred to as the Leray Limestone with overlying shaly beds representing the Glenburnie (see figure 6). Within the Oak Hall-Centre Hall succession, recognition of the three K-bentonites (N1, N2, and N3), within an interval that grades from massive calcarenite facies of the Oak Hall into shalier, bioturbated Centre Hall is similar to the conditions found in the vicinity of the Hounsfield K-bentonite within the Leray to Watertown interval. This is an important consideration since the Hounsfield K-bentonite in New York (sensu Kay, 1931) occurs just below the base of the Watertown and appears to be the equivalent of the Millbrig (Mitchell et al., 2004; see further discussion elsewhere herein). If either of the N1 or N2 K-bentonites are the Millbrig, Kay's assertions that the Oak Hall is Rockland and the Centre Hall and Rodman are Kirkfield (as shown on figure 10) are certainly in error.

With regard to the recognition of unconformities, the substantial set of data presented along with data considered herein suggests that both Kay and Rones were at least partially correct. There may be two unconformities one at the base of the Centre Hall and one at the base of the Oak Hall, a pattern that also appears to be true in New York. However, based on correlation of this interval with the type region, it is proposed herein that the true position of the Black River – Trenton group contact actually lies above both the Linden Hall and Oak Hall, and likely lies near the contact of the Centre Hall and the Rodman Member and perhaps at the position of the N3 K-bentonite. Thus although future emphasis should be placed on redefinition of the Black River – Trenton Group contact in Pennsylvania, for consistency with stratigraphic nomenclature, the contact of the Black River Group is placed at the top of the Linden Hall and base of the Centre Hall following the use of Rones (1955, 1969). Thus the Nealmont Formation as currently defined is still considered to be part of the Trenton Group.

Distribution

As defined and recognized in the subsurface, the Black River Group is recognized over a substantial area of eastern North America and with minor variation is characteristically easy to recognize both in the subsurface and in outcrop for the dove tan-to-light-grey, micrite-dominated limestones (Keith, 1988). In the Cumberland Valley the equivalent of the Black River Snyder and Linden Hall Formations is the Chambersburg Group. Stose (1906, 1909) named the Chambersburg Limestone for the cobbly-weathering dark crystalline limestones and shaly argillaceous limestones that sit between the Stones River Limestone below and the Martinsburg Shale above. With exclusion of the basal calcilutites of the St. Paul Group (by Butts, 1940), the unit ranges upward to a maximum of about 230 meters in the vicinity of Chambersburg (slightly north of the depocenter for the underlying beds). Craig (1941) divided the Chambersburg Limestone into three formations including the Shippensburg, the Mercersburg, and Greencastle Formations (**see figure 5**). The Chambersburg was also correlated southward into West Virginia (Cardwell et al., 1968), Maryland, and the Shenandoah Valley of Virginia, although Cooper and Cooper (1946) introduced another nomenclature for this area and the unit as a whole was substantially thinner (about 69 meters) especially in Maryland (Nutter, 1973).

Lithologically, the Chambersburg contains a narrower range of lithofacies compared to its equivalents in the northern portions of the Ridge and Valley. As suggested by Bassler (1919) and Kay (1944) the Chambersburg depositional region may have been somewhat isolated from the latter region by a pronounced barrier with the Chambersburg deposited in a more open-marine and potentially deeper-water setting. Root (1968) described the Chambersburg as a dark gray, argillaceous, cobbly-weathering limestone with abundant shaly partings. He also noted the presence of a number of K-bentonites that likely correspond to those in the Linden Hall to lower Nealmont, but are as yet unstudied and undiagnosed.

Biostratigraphic assessments and recognition of at least five different faunal zones within the Chambersburg indicate that the unit is equivalent to the Black River Group (Bassler, 1919). Since the early work of Bassler very little research has been done to assess biostratigraphy of the Chambersburg despite its important echinoderm, brachiopod, and trilobite faunas. Nonetheless, Swain (1957) identified the boundary between his *Aparchites kaufmanensis* and *Monoceratella teres* ostracod zones just below the base of the Chambersburg Limestone where it rests on the New Market Formation of the St. Paul Group. In West Virginia, Maryland, and northern Virginia, work on conodonts recognized the Chambersburg as containing conodonts from the *Baltoniodus gerdae* subzone of the long-ranging *Amorphognathus tvaerensis* biozone (Harris and colleagues, 1994). Thus, as defined, the base of the Turinian is coincident with the base of the Chambersburg Limestone in the Cumberland Valley region.

Unlike the underlying Chazy Group and equivalents, the Black River Group has a much more constant thickness except for regions in the vicinities of some major topographic highs that were evidently exposed during the deposition of the Black River. Such topographic highs are documented in the vicinity of the Adirondack Arch, the Canadian Shield of New York, and possibly even in eastern Pennsylvania along the New Jersey Arch as identified by Read (1989). As mentioned by Craig (1941), the Chambersburg is not recognized beyond the Lebanon Valley (he states that it is not present east of the Susquehanna River), although an uppermost Chambersburg and basal Martinsburg equivalent unit is recognized (Jacksonburg Limestone). This would suggest that this region was uplifted and not influenced by deposition during the majority of the Turinian and only in the late Turinian to early Rocklandian did sedimentation resume.

Snyder Formation

Defined as the basal member of the Benner Formation by Kay (1944), and subsequently raised to formation status by Rones (1969), the Snyder was considered to be the equivalent of the lower Pamelia Formation of the type Black River Group. At the type section at Union Furnace, Kay defined an interval of nearly 30 meters of light-grey weathering relatively pure limestones and dolomitic limestones representative of his Snyder (**see Figure 7**). On fresh surfaces the limestone is a medium-to-dark grey to greenish-grey and varies from burrow-mottled mudstones and skeletal wackestones to packstones and bioclastic grainstones. In addition, the unit is recognized to contain significant intervals of intraformational (and extraformational?) conglomerates, intervals of cross-bedded grainstones, ooids, and a few K-bentonite layers. Rones (1969) recognized five main lithologic units within the Snyder (lettered A to H in stratigraphic order). These units are as follows: Unit A was termed the oolitic calcarenite beds; Unit B the lower laminated, "Dolomitic" calcisiltite to fine-grained grainstone beds; Unit C the lower bioclastic and "mud-pellet" calcarenite beds with minor *Streptelasma* and *Tetradium*-rich calcilutites; Unit E-D was referred to as the chemical lime beds and was dominated by light-gray fine calcilutite facies grading upward into *Tetradium*-rich beds; Unit F was the upper

“Dolomitic” which was similar to Unit B; Unit G was referred to as the mud-cracked beds and contained birdseye and *Skolithos* or *Phytopsis*-style burrows, fenestral fabrics and an interval of mud-cracked lutites interbedded with thin black shales; the uppermost unit of the Snyder, Unit H, was referred to as the upper bioclastic beds and was represented by massive-megaclastic conglomerates, fossiliferous calcarenites, with numerous biostromal wackestones, algal (*Solenopora*/*Girvanella*) rudstones, and was capped by a final ooid-grainstone unit. As suggested by Berkheiser & Lollis (1986) the last unit contains two K-bentonites which they informally referred to as B1-A and B1-B. Recent field work has produced Bathyurid trilobites from the vicinity of these K-bentonites.

As pointed out by Kay (1944) and supported by Rones (1969), the conglomerate and ooid-rich intervals of this unit are easily recognized across much of the region and enable it to be widely correlated without difficulty. It stands in stark contrast to the underlying Hatter Formation in that it contains substantially less siliciclastic material. Only the basal unit (Unit A of Rones, 1969 or mining unit 8 of Berkheiser & Lollis, 1986) contains appreciable insoluble residues (up to 20.8 percent). In contrast, the remainder of the Snyder is characterized by insoluble residues ranging between 2.6 to 8.6 percent (Berkheiser & Lollis, 1986). Due to its relatively pure limestone composition (and the occurrence of the chemical lime beds) the unit has often been quarried and exposures have been accessible for investigation. In the western Ridge and Valley through the Kishacoquillas and Path Valleys the Snyder is easily recognized although it becomes truncated from the top with the uppermost units recognized by Rones removed at the base of the Nealmont Formation especially in Path Valley. An equivalent intraclastic unit is also present at the top of the Shippensburg Limestone in the Doylesburg Member where the unit is dominated by fine-grained carbonates (calcilutites) developed into an interval of heavy limestone conglomerate beds typically dominated by *Nidulites pyriformis* of early workers, now recognized to be a Dasycladacean algae (Osgood & Fischer, 1960). Thus across the region, the Snyder varies from a maximum of just over 33 meters in the western valleys with an average of about 21 meters. Note this is in contrast to the value established by Faill and others (1989) for the thickness exposed at Union Furnace, which was established at 55 meters as they included all strata up to the uppermost ooid-bearing conglomerate. Hence the 55 meters doubled the thickness recognized by all other workers including that identified by Berkheiser and Lollis (1986). Given their values, these authors included the entire interval up to the base of the Nealmont and comprising the bentonite-rich interval recognized to be the Valley View, thus these values are in error. Southeastward of the Nittany Valley region, the thickness of the Snyder is only about 15 meters in the Black Log Valley (at Orbisonia), and is recognized to a low of about 6 meters in the Path Valley where it is overlain by the Nealmont Formation (Kay, 1944, Rones, 1969). In the Cumberland Valley region, the thickness of the Snyder-equivalent the Doylesburg ranges from less than two meters to about 12 meters in the Franklin County region (Craig, 1949).

In the subsurface Wagner (1966) reported the base of the Snyder to be unconformable as the basal units recognized by Rones (1955, 1969) appear to “pinch and swell” against the top of the Hatter. The top of the Snyder is placed below the lowest microcrystalline limestone or fine-grained calcilutites of the Stover Member of the Linden Hall which are usually underlain by the coarser-grained Snyder facies with sharp contact although it is described as conformable (Rones, 1955, Wagner, 1966). North and south of the Ridge and Valley in southern New York and West Virginia, Wagner reported that the beds below the calcilutites of the Stover often contain thin beds of dolomitic limestone (as reported in outcrop by Rones). These dolomitic intervals are characteristic of the Pamelia Formation of New York, although Wagner (1966) favored correlation of the Snyder with the upper Lowville and basal Chaumont of New York, based on his correlation of the underlying Hatter Formation with the Pamelia and despite Kay's (1944) support for the Snyder as lower Pamelia.

Biostratigraphically, the Snyder Formation contains a fairly restricted association of coral-rich faunas. These include *Tetradium cellulosum*, large favositid corals resembling *Foerstephyllum halli* as well as a species of rugose coral – likely either *Lambeophyllum profundum* or *Streptelasma corniculum*. These coarse-grained intervals also contain minor echinoderm material as well as small brachiopods. In the middle zone, Units E-D and F, the faunas are characteristically more depauperate and are limited to a few ostracods, *Girvanella* algae, a range of small twig bryozoans, and gastropods. East of the Path Valley, the Snyder equivalent becomes more fossiliferous and contains a greater faunal diversity thus suggesting the presence again of a significant restrictive barrier that divided the central and western Valley and Ridge area from the eastern Valley & Ridge (Great Valley region).

Linden Hall Formation

The Linden Hall Formation was proposed by Rones (1969) to replace the interval originally conceived of as the Benner and Curtin Limestone formations by Kay (1944). At the type section of the Linden Hall east of State College (the Oak Hall Quarry), the Linden Hall is a 46 meter-thick interval of characteristically thick-to-massive bedded limestones underlying the irregularly bedded Centre Hall Member of the Nealmont Formation and overlying the Snyder Limestone (Rones, 1969, **see figure 9**). Overall the Linden Hall Formation is dominated by greenish-grey to brownish grey, structureless micritic limemudstones, bioturbated wackestones, and occasional packstones. The lower unit is extensively bioturbated and shows wavy laminae, occasional intraclastic conglomerates, a number of hardground surfaces, cherty horizons, and K-bentonites. The upper units are typically finer-grained massive micrites. Berkheiser and Lollis, (1986) identified a number of closely-spaced, and relatively thick K-bentonites in the Linden Hall at Union Furnace that correspond to those recognized by early workers including Rosenkrans (1932).

Faill and colleagues (1977, 1989) during mapping of different areas of the Ridge and Valley suggested that the Linden Hall should include all units between the Snyder and the Rodman member of the Nealmont. However this usage appears not to have been accepted and herein the usage generally follows that of Rones (1969), although Berg and colleagues (1980) referred to the interval as the Benner after Kay's original term. Overall the Linden Hall as currently defined contains four main units or members which include the Stover, Valley View, Valentine, and Oak Hall (**figure 10**).

Stover Member

The Stover Limestone has its type section at Union Furnace in Blair County where it is currently recognized to be about 16.5 meters thick, as opposed to 27 meters as originally conceived by Kay (1944). The uppermost 10.5 meters of Kay's Stover has been removed and included in the overlying Valley View Member owing to recognition of the numerous K-bentonites immediately overlying the 16.5 meter succession, which were defined elsewhere to occur within the Valley View Member (**figure 12**). Originally, the Stover formed the upper member of Kay's Benner Formation and was defined as the very dark grey, heavy ledged, fine-grained limestone with thin stylolitic partings (wavy lamination) and fucoidal mottling

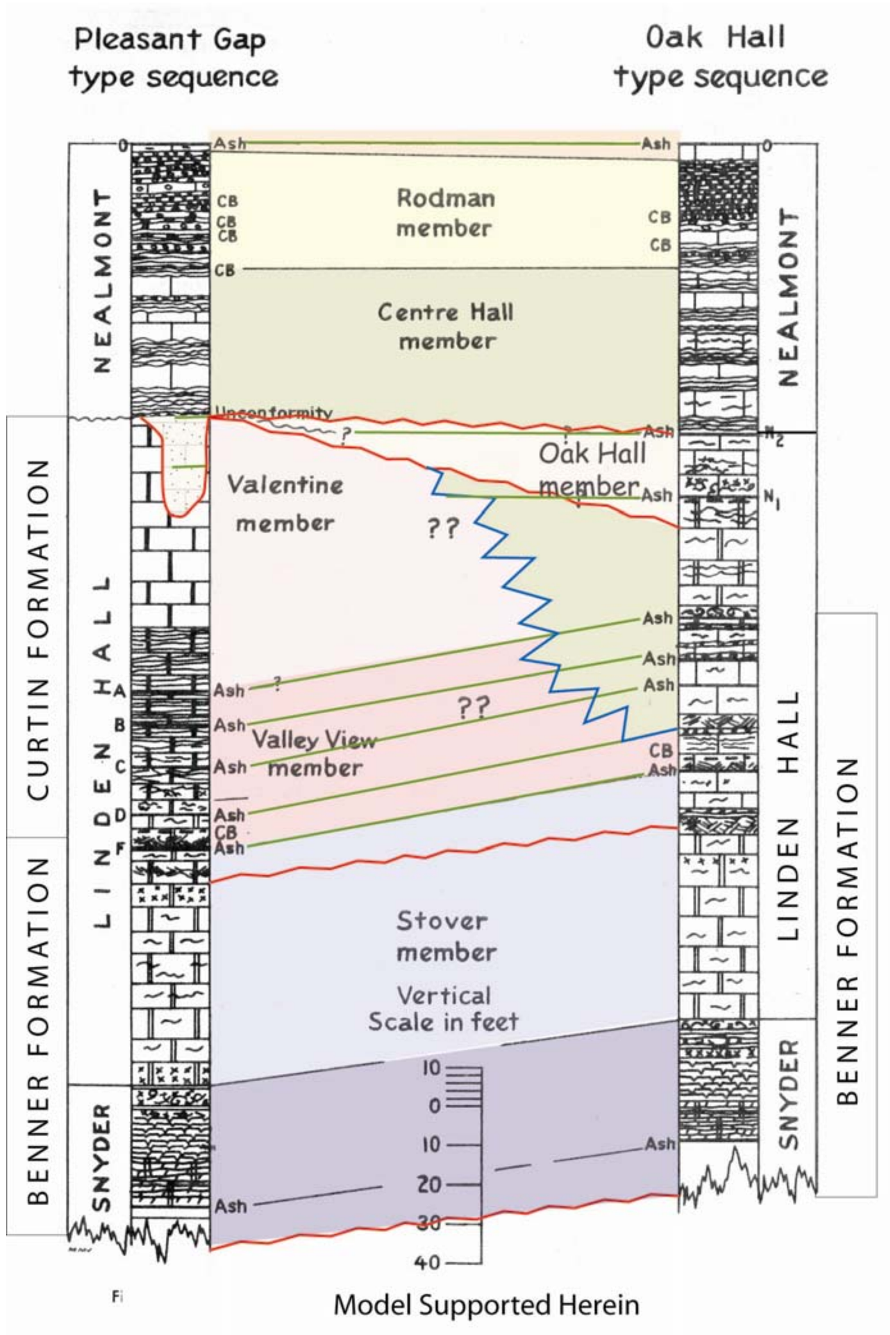


Figure 11: Stratigraphic model for the Upper Black River Group in the southeastern Nittany Valley region based on data from both Kay (1944) and Rones (1969) and observations herein. Model is to be compared to the stratigraphic models for the interval by Kay and Rones as shown in Figure 9. Key change is recognition of a facies lateral equivalent of the Valley View member following the suggestions of Rones (lower beige). Nonetheless, the Oak Hall member is restrained to the original concept of Kay, which results in the distinction of the lower Oak Hall of Rones as a separate unit that sits below the unconformity recognized by Kay (1944). Also shown are other potential discontinuities within the succession.

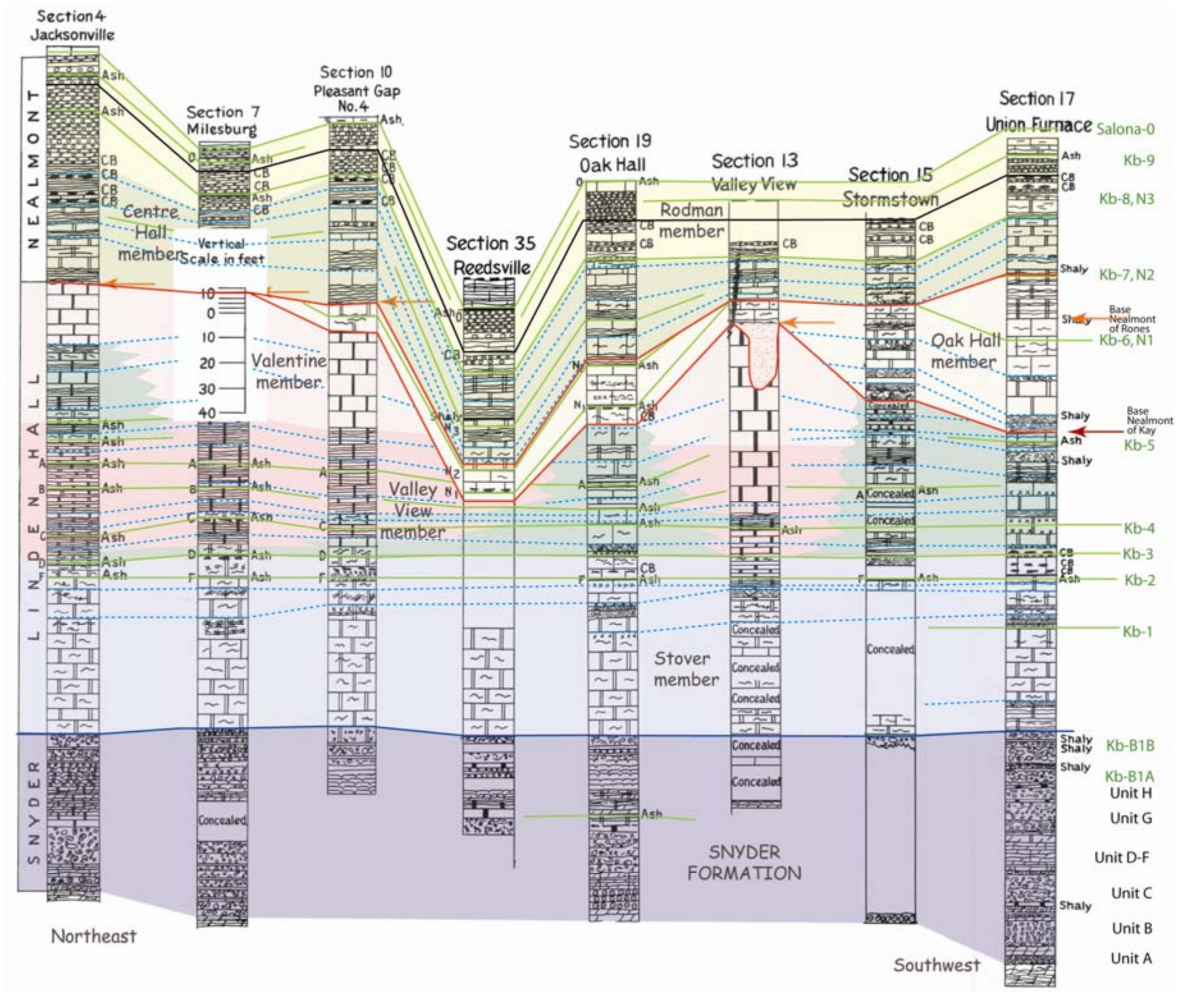


Figure 12: Northeast-southwest oriented correlated cross-section of the for the western Ridge and Valley (Jacksonville, Milesburg, Pleasant Gap, Oak Hall, Valley View, Stormstown, Union Furnace), and the section at Kishacoquillas Valley (Reedsville). Jacksonville is 66 km from Union Furnace along strike, while the distance from Pleasant Gap to Reedsville is 29 km. Marker horizons including K-bentonites, small-scale cycles, and specific lithofacies have been used in correlation. K-bentonites abbreviated (for example: Kb-1) are after those recognized by Berkheiser & Lollis (1986), and units in the Snyder are after Rones (1969). Datum plane is established at the level of the F K-bentonite.

(bioturbation). Rones (1969) identified the wavy lamination as partially dolomitized clay-rich seams with occasional quartz grains. This style of facies weathers to medium-grey to brownish grey in color and is dominantly a fine-grained calcilutite to bioturbated wackestone lithology. This contrasts strikingly with the underlying Snyder Formation which is substantially more coarse-grained and conglomeratic. The upper Stover becomes somewhat more dolomitic near its top. At Union Furnace the basal contact was placed at the top of Unit H (of Rones, 1969) at the level of an ooid-grainstone which is locally developed into a fossiliferous intraclastic conglomerate (Kay, 1944). The upper contact of the Stover was placed by Kay at the position of Rosenkrans' "e K-bentonite." Rones (1969) didn't recognize the e K-bentonite at Union Furnace and established the contact at the position of the distinct d K-bentonite (which is underlain by a distinct chert bed) as shown in **figure 12**. This latter K-bentonite corresponds to the B-3 K-bentonite of Berkheiser and Lollis (1986) at Union Furnace (see figure 7).

The Stover Member has been widely recognized southward along the western Ridge and Valley and has been extended into Maryland, West Virginia and even into Virginia by previous workers (Twenhofel et al., 1954, Kay 1956). Nonetheless, the Stover was renamed the McGraw Limestone in Virginia by Kay (1956). Throughout this region, the Stover maintains a remarkably even thickness. Eastward, the Stover eventually becomes substantially truncated along the Adirondack Axis and is completely truncated beneath the Nealmont Limestone and is not recognized at Orbisonia in the Black Log Valley. Further east, an equivalent unit is recognized in the Cumberland Valley where the Stover is correlated with the Housum Member of the Chambersburg Limestone. Although bed-specific correlations are lacking the Housum has been noted to contain a K-bentonite near the top of the unit that may correlate with K-bentonite F of the Stover (which occurs below the e K-bentonite of Kay) (Craig, 1949). In the Cumberland Valley the Housum is an interbedded crinkly-weathering, dark-gray limestone with numerous argillaceous wavy partings that appear similar to the Stover, but are shalier and somewhat more fossiliferous. These beds often weather to a light gray cobbly appearance and are recognized to be about twenty meters-thick at the type area southwest of Chambersburg.

In the subsurface, Wagner (1966) noted the presence of the Benner Formation although he didn't differentiate specific components except to indicate that in southern New York, the entire lower to middle Benner (including the Valley View interval) was dolomitized locally to a coarsely crystalline dolostone. This specific stratigraphic interval may be one of the intervals responsible for producing significant quantities of natural gas in the region today. Although gas production appears to be related to structural trends, the relatively fine-grained, low porosity facies of the Stover located immediately above the relatively coarse-grained facies of the Snyder may represent an important stratigraphic association for subsequent dolomitization and trap development.

Biostratigraphically, the Stover is recognized for the abundance of *Camarocladia* sp. (*gracilis*?), which was considered to be a sponge (Sinclair, 1956). The interval was also known as the *Cryptophragmus antiquatus* beds based on the abundance of these organisms (Kay, 1944) which may be a sponge as well. The Stover has also been known to produce a number of other sponge-like forms including *Distactospongia* sp. and *Stromatocerium* sp. Kay (1944) also reported the echinoderms *Pleurocystites* sp. and *Dendrocrinus* sp., at least six different brachiopod genera, several large cephalopods, several genera of trilobites as well as *Maclurites bigsbyi,* which is prevalent in the Witten Limestone of Virginia. In contrast to the underlying Snyder, this interval contains significantly fewer coral taxa, but still contains *Tetradium*. To the southeast in the Great Valley region, the lower member of the Mercersburg Formation (i.e., the Housum Member) contains a very diverse fauna as reported by Bassler (1919) from the northern Maryland to southern Pennsylvania portions of the Cumberland Valley. In this region, the fauna is characterized by an echinoderm-rich fauna from the top of the *Nidulites* (receptaculitid) interval. It contains genera related to *Amygdalocystites, Pleurocystites*, *Bolboporites, Carabocrinus,* and *Porocrinus*; at least six genera of bryozoans including *Prasopora contigua*; 11 genera of brachiopods including two forms of *Sowerbyella* one of which may be *S. lebanonensis*; as well as a number of trilobites including *Ceraurus pleurexanthemus*. These faunas suggest an affinity to the Lebanon Limestone of Tennessee with which it is correlated, and according to Kay (1944) the occurrence of the *Cryptophragmus* and *Distactospongia* suggested a correlation with the Pamelia of New York and Ontario – a position that is also supported herein.

Valley View Member

The Valley View Member and the overlying Valentine Member are two of the most economically important limestone facies of the entire Black River Group. These rocks are the "chemical limes" of early workers and are the combined equivalent of what was originally named the Valentine Limestone by Field (1919). The type section for the revised Valley View was established in the quarries and mines north of Bellefonte near the vicinity of Valley View by Kay (1943). In contrast to the overlying Valentine, the Valley View Limestone is a thin-bedded micritic limestone unit described as relatively impure due to the occurrence of the zone of abundant K-bentonites (K-bentonites A through D of Rosenkrans, 1932) and cherts, with some disseminated quartz grains of detrital and/or volcanic origin within intervening strata. At Bellefonte, Kay (1944) described the Valley View as a light-gray, white-weathering laminated fine-grained limestone with argillaceous partings and interbedded with thicker shales and somewhat coarser limestones. The base of the unit was apparently more bioturbated in similar fashion to the underlying Stover Member with substantially more argillaceous beds toward the top where the K-bentonites are more closely spaced. Berkheiser and Lollis (1986) report this interval to contain the highest insoluble residue proportion of the entire Linden Hall Formation with values ranging between 5.2 and 7.0 percent which is higher than the average of ~ 4.0 percent or less.

The Valley View and Valentine Members were collectively mapped from the Nippenose Valley southwestward to Pleasant Gap and Bellefonte and were included in the Curtin Limestone by Kay (1944). These facies reach a maximum thickness of about 45 meters to the northwestern edge of the Ridge and Valley. In the subsurface, Wagner (1966) did not differentiate them from the underlying Stover member. However, in outcrops to the south and east including the quarry at Tyrone, virtually no lithologies of the Valley View or Valentine were recognized, nor were these specific facies located at Union Furnace (Kay, 1944). However, as mentioned, K-bentonites A-D are now recognized at Union Furnace, and although the facies within which they occur is somewhat coarser than at Bellefonte and characteristically more bioturbated in similar fashion to the underlying Stover Limestone. Thus, the top 10.5 meters of Kay's Stover are now considered to be equivalents of the Valley View Member although they are more bioturbated and less laminated except in the upper portion of the unit (**see figure 11**). To the east of State College at Oak Hall, these lithologies were also not uniformly developed, the shaly, ribbon lutites of the Valley View become somewhat more bioturbated and slightly more coarse-grained with interstitial sparite (birdseyes?; Rones, 1969). These facies are very similar to those at Union Furnace which Kay included in the Stover.

This facies change was suggested by Rones (1969) to represent a change into what Kay (1944) defined as the Oak Hall Formation, although it is clear that Rones' use of Oak Hall was in an expanded sense compared to that of Kay (1943, 1944). Previously, the lack of Valentine and Valley View facies at Oak Hall and at Union Furnace had been used as partial evidence to suggest the presence of the Black River-Trenton unconformity (as discussed earlier). However herein, the stratigraphic concepts for each unit have been re-evaluated resulting in the correlations shown in **figures 11 and 12** which are to be compared to the earlier models (**figure**

9). The key resolutions that are suggested include: 1) lowering the contact of the Stover Member (as per Rones, 1969), 2) correlation and recognition of facies equivalents of the Valley View Member as suggested by Rones (1969), 3) retaining the distinction of the Oak Hall Member bracketed below by unconformity as suggested by Kay (1944), 4) retaining the distinction of the top Oak Hall unconformity as recognized by Rones, 5) separation of the lower Oak Hall of Rones as an unnamed facies of the Valley View-Valentine, and 6) retention of the Oak Hall (of Kay, 1944) as the upper and youngest member of the Linden Hall These adjustments are supported by K-bentonite correlations and recognition of depositional trends observed in the correlated cross-section.

Thus, as defined herein, the Valley View is now extended in its distribution with lateral equivalents recognized although facies become, slightly more coarse-grained, more fossiliferous and similar to the underlying Stover. Nonetheless, the Valley View and its lateral facies equivalents maintain maximum thickness in the northeastern Nittany Valley and thin somewhat to the south with thickness in outcrops ranging from about 10 to 15 meters. It is easily recognized in outcrop owing to its relatively whitish-cream colored weathering and the abundance of yellowish green unctuous clay seams (K-Bentonites) that weather recessively. Outside of the northwestern Ridge and Valley the Valley View interval is more difficult to recognize, although a bentonite-rich interval has been noted in the Mercersburg Formation (Kauffman member). Facies of the Kauffman Member, as reported by Craig (1949), include dark-gray, fine-to-medium-grained, micritic wackestones that weather to thin, crinkly beds. The facies becomes more cobbly to the northeast, but are interbedded with thin to medium-bedded quartzitic siltstones to the south toward the Maryland border. The presence of four K-bentonites in the sections south of Chambersburg suggest correlation with the Valley View Member. In this region, the maximum thickness is about 50 meters. In northern Virginia, the Kauffman Member transitions into the Eggleston Limestone which contains the Deicke and Millbrig and at its base, the Walker Mountain Sandstone, which may correlate with the basal Kauffman siltstones.

As with the overlying Valentine, the Valley View Member is sparsely fossiliferous and was deposited in an environment that was fairly restricted from normal marine circulation. Where faunas are found, the unit is characterized by small colonies of the tabulate coral *Tetradium cellulosum*, *Thamnobeatricia parallela* (with affinities to the labechiid stromatoporoids), a number of trilobite taxa including *Bathyurus extans*, and occasional cephalopods, bivalves, and ostracods, which are all more abundant in the shalier interbeds.

Valentine Member

The upper member of Kay's (1943, 1944) now abandoned Curtin Formation, or the third member of Rones' Linden Hall Formation (sensu, 1969) is the Valentine Limestone. The unit was first proposed by Field (1919) as the "pure quarry rock" of the Bellefonte region of Centre County. Kay (1943) removed the lower Valentine of Field (1919) to form the Valley View Member, leaving the upper interval as a distinct facies. Berg (1980) includes the unit in the Benner Formation which is used in place of the Linden Hall, although it appears that both terms are in use by the Pennsylvania Geological Survey (i.e. see Laughrey et al., 2004). The Valentine is a characteristically massive micritic limestone (calcilutite) that is nearly structureless except for development of several hardground style contacts and a somewhat laminated lower six meters (Rones, 1969). It is characteristically light-grey to white and massive bedded. Although it varies in thickness, it is approximately twenty meters thick at Bellefonte (Kay, 1943). In hand-sample, the conchoidal fracturing calcilutites show disseminated calcite spar attributed to fenestral fabrics, but the unit lacks evidence for vertical burrows (*Phytopsis)* that are often associated with these features. In some cases, the larger spar-filled lenses have been attributed to broken masses of *Tetradium* especially where they are characteristically angular in cross-section. In thin-section, this unit appears to be composed of small peloids that have a somewhat clotted texture with internal filamentous structure, suggesting perhaps algal origin.

In outcrop sections to the south and southeast (i.e. Oak Hall, Reedsville, Stormstown, Union Furnace), the Valentine Member (and the underlying Valley View Member) appears to transition into a bioturbated wackestone that is distinctly more fossiliferous and only minimally coarse grained. It is typically developed into a Stover-like facies and is often difficult to distinguish from it. Small-scale shallowing upwards cycles become more evident in these facies and can be used in correlation between closely spaced outcrops as shown in figure **12**. In these regions, the K-bentonites are usually preserved and together with the shaly-based cycles can be used to correlate between sections. This particular facies, although present at Oak Hall, is not the same facies as was described as the Oak Hall Member by Kay (1944), which is evidently an even more coarse-grained, cross-bedded unit that can be found incising underlying units. Thus the “Oak Hall facies” of Rones is herein considered a lateral equivalent of the Valley View and Valentine and is not the equivalent facies designated as Oak Hall by Kay. Usage herein follows that of Kay, although the unit is retained in the Linden Hall Formation as discussed earlier.

The base of the Valentine is drawn at the position of the uppermost K-bentonite recognized, which is typically the “a K-bentonite” of Rosenkrans (1934). At Jacksonville, Rones identified two additional K-bentonites above the “a”, the uppermost of which corresponds to a K-bentonite labeled B5 by Berkheiser and Lollis (1986) at Union Furnace. Where the “a K-bentonite” is not specifically recognized the contact is placed at about the level of the first massive calcilutite above the Valley View. The top of the Valentine at the type region is placed at the base of the Centre Hall Member of the Nealmont Formation or at the base of the Oak Hall Member of the Linden Hall Formation. The stratigraphic relationship of this contact is complex as it is traced both southwestward in the Ridge and Valley and to the southeast into the central Ridge and Valley as at Reedsville as shown by the red line in **figure 12**. Beginning in the south and progressing northeastward, the Valentine is onlapped first by the Oak Hall Member of the Linden Hall and then by the Centre Hall Member of the overlying Nealmont Formation.

Thus, across its exposure area, the unit ranges in thickness from zero to a maximum of about 27 meters in the mine near Valley View. In this location the lower half is probably an equivalent of the Valley View, although most of the K-bentonites are missing and only one or two occur in the lower three meters. It is more typically about 12-15 meters thick. As mentioned the unit is nearly structureless, and bedding is only evident through development of partings and hardground surfaces that occasionally show evidence of borings and minor evidence of cryptalgal lamination. Southeastward, the Valentine appears to be completely truncated at Reedsville in the Kishacoquillas Valley and is also not recognized at Orbisonia in the Black Log Valley. With recognition of a possible karstic contact at the top contact of the Valentine, it appears that the region where the unit was deposited was cratonward of a developing topographic high that evidently became a prominent feature during deposition of the Valley View Member and Valentine Member when facies became more heterogeneous within the region. This structure was referred to as the Adirondack Arch by Kay (1948) which likely developed due to late stages of the Blountian Tectophase leading into the Vermontian Tectophase of the Taconic Orogeny. **Figure 13** is the same cross-section shown in figure 12, but is drawn with a datum plane at the

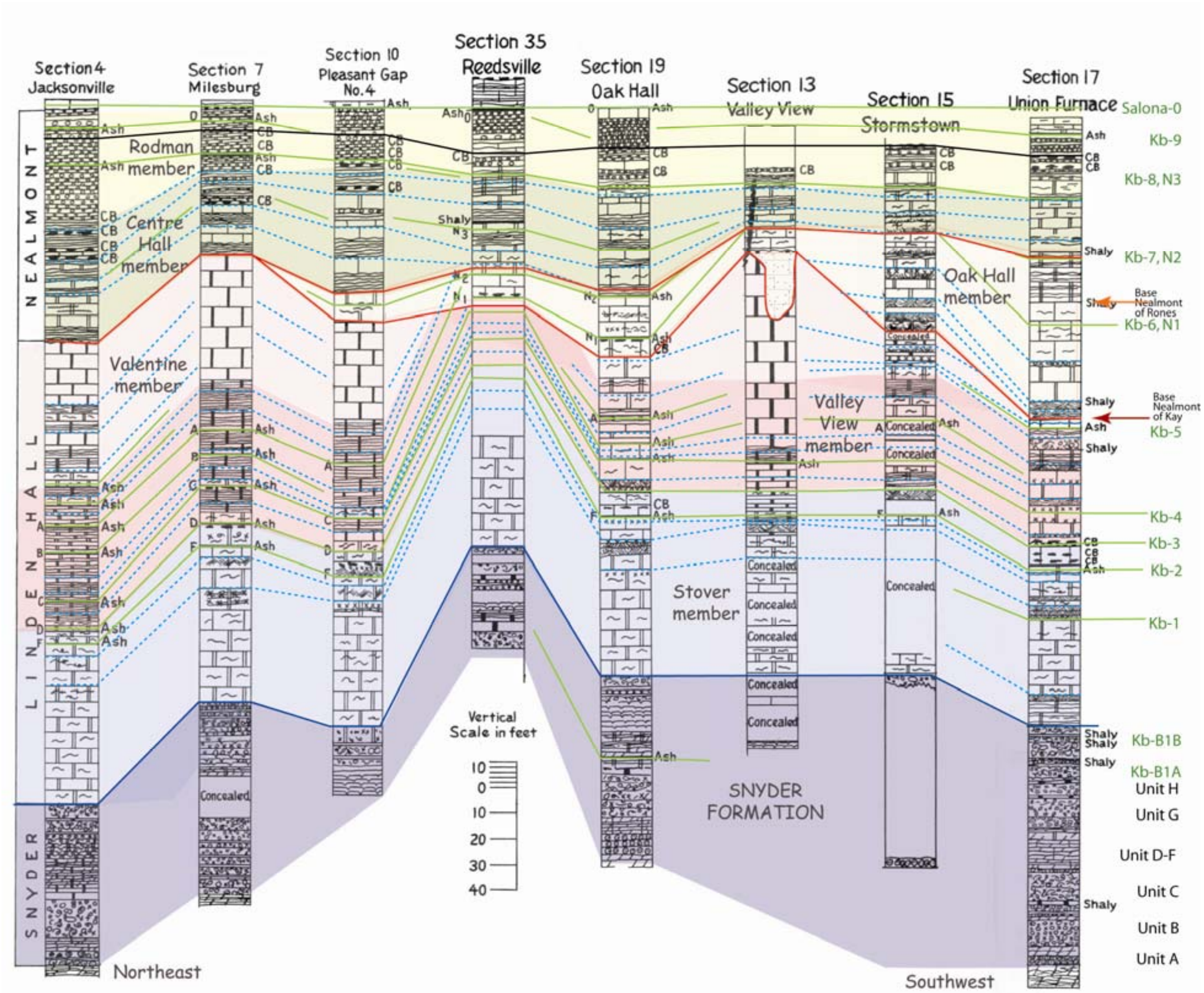


Figure 13: Northeast-southwest oriented correlated cross-section for the western Ridge and Valley (Jacksonville, Milesburg, Pleasant Gap, Oak Hall, Valley View, Stormstown, Union Furnace), and the section at Kishacoquillas Valley (Reedsville). Jacksonville is 66 km from Union Furnace along strike, while the distance from Pleasant Gap to Reedsville is approximately 29 km. Marker horizons including K-bentonites, small-scale cycles, and specific lithofacies have been used in correlation. K-bentonites abbreviated (for example: Kb-1) are after those recognized by Berkheiser & Lollis (1986), and units in the Snyder are after Rones (1969). In contrast to Figure 12, the datum plane is established at the level of the Salona 0 K-bentonite that is roughly equivalent to the contact between the Nealmont Formation and the overlying Salona Formation.

position of the Salona 0 K-bentonite which roughly forms the contact between the Nealmont and the Salona Formations. As shown in the diagram, sections to the south and southeast show truncation of the Valley View-Valentine interval against a newly developed topographic high.

Oak Hall Member

The final member of the Linden Hall Formation is the Oak Hall Member. Originally defined as the basal member of the Nealmont Formation, Kay (1944) recognized a one meter-thick, impure, coarse-grained limestone that formed the base of his Nealmont at Union Furnace (**see figures 7 & 13**). This coarse-grained limestone was recognized as a fairly massive ledge of cross-bedded calcisiltite (calcarenite?) by Rones (1969) that sat immediately above somewhat coarse-grained but argillaceous, bioturbated facies, which were thought to be uppermost Stover by Kay (1944) and undifferentiated Linden Hall by Rones (1969). Given the number of K-bentonites in the interval underlying the coarse-grained ledge– the interval is considered a lateral equivalent of the Valley View Member of the Linden Hall, with very little evidence for Valentine-type lithologies. Thus, the base of the massive ledge forms an unconformity between the Oak Hall and the underlying limestones, and regionally appears to truncate into the lowest Valentine to uppermost Valley View interval by erosional (karstic?) unconformity (Kay 1944 – also see discussion elsewhere).

At Union Furnace, the basal Oak Hall grades upward into an interval of medium-to-coarse grained limestone facies for a total of approximately 17 meters, the uppermost of which becomes more shaly in the transition into the overlying Centre Hall. Kay recognized two K-bentonites in this interval which were called the N1 and the N2. The N1 was located approximately five meters below the top and the N2 was used to help define the contact with the overlying Centre Hall. Rones (1969) described the lithology of this unit as a thick-bedded, dark gray, medium-to-coarse limestone with scattered irregular dolomitic bands that appear in wavy ribbons much like those in the Stover Limestone. These become interbedded with brachiopod packstones and the unit can show coarser cross-bedded intervals and intraclastic beds. Rones also noted the prominent K-bentonites, and noted that the Oak Hall was dominated by small limestone clasts imbedded in a matrix of micrite and noted the presence of euhedral dolomite rhombs and euhedral quartz grains.

At Oak Hall, the type region selected for the unit by Kay, the coarse grainstone ledge described by Kay is less pronounced and the shaly interval at the top of the Valley View is less evident, although at least 4 of the 5 K-bentonites have been recognized (Rones, 1969). Nonetheless the N1 and N2 K-bentonites are well developed and occur in a zone of cross-bedded calcarenites. Below this seven meter thick interval the beds are mostly developed as calcisiltites with brachiopod stringers. Rones (1969) divided this section into an upper and lower part respectively, with the lower part having a substantially purer composition with less than two percent of the unit containing insoluble residues. Combined with the subjacent K-bentonites, these low values suggest equivalency of the unit to the Valley View and Valentine. In contrast, the upper portion of Rones' Oak Hall had up to 7% insoluble residues (and higher in some levels), most of which was a silica fraction which Rones felt was derived from the N1 and N2 K-bentonites although a minor amount may also be detrital as some quartz appeared to be fairly rounded. The broadening of silica peaks into overlying and underlying strata surrounding K-bentonites was attributed by Rones to be a function of larger grain-size, increased porosity, and increased bioturbation in this unit, which was in contrast to the peaks in the Valley View that were very narrow and pronounced at the level of the K-bentonite.

Further north of Oak Hall, at Pleasant Gap and at Bellefonte, Oak Hall lithologies are characteristically thin and are recognized as relatively thin medium-grained calcisiltites with brachiopod packstone stringers. As noted by Kay (1944) the Oak Hall was limited to a thin interval that showed definite evidence for channeling into the underlying Valentine and described the coarse-grained facies in some detail as reported earlier. At Pleasant Gap, the coarse facies is recognized intermittently, but quarrying activities do produce exposures that show development of the N1 and N2 K-bentonites. In this location the Oak Hall is just over three meters thick and further north at Milesburg and certainly at Salona near Jacksonville no coarse Oak Hall facies are recognized (Rones, 1969). At the Ashcom Quarry in Bedford County in the southwestern Ridge and Valley, the succession at the base of the Nealmont appears to be more similar to the Oak Hall facies although it is somewhat thicker and is less distinct from the overlying Centre Hall that has become more nodular in appearance. Given this distribution pattern, the Oak Hall appears itself either to be capped by an unconformity (as promoted by Rones) or represents an onlap unit from the south that is floored by an unconformity and capped by a flooding surface that helps preserve the K-bentonites that enable its correlation.

Faunally, the Oak Hall member records the re-introduction of a number of taxa and witnesses the invasion of new forms, a pattern that continues upward into the overlying Nealmont Formation and reverses a trend toward reduction of taxa in the underlying Valley View and Valentine Members. Similar to the Snyder, the Oak Hall is dominated by a coral-stromatoporoid assemblage. Although most fossils are poorly preserved, a number of cnidarian forms are recognized. As reported by Kay (1944), these include *Columnaria halli, Lambeophyllum profundum,* and *Paleoalveolites paquettensis*- most all of which are found in the Watertown to lower Bobcaygeon interval of New York which led Kay to favor an upper Black River to lower Rocklandian age for this unit. Other forms in the Oak Hall include *Sowerbyella punctostriatus*- which is equivalent to *Sowerbyella curdsvillensis* var. *plana* as suggested by Titus (1984), and *Leptaena charlottae.* The latter has also been recognized from the upper Chambersburg Limestone in the Cumberland Valley as well as from the Decorah Shale in Minnesota (Bassler, 1919). A number of trilobite taxa are reported as well. The other main taxon recognized from this interval is *Maclurites logani* which again is characteristic of the Watertown to lower Rockland interval of New York and Ontario (Kay, 1935, 1944).

Trenton Group: General Description, Contacts & Distribution

In the Ridge and Valley of central Pennsylvania, black Trenton argillaceous limestones have been recognized on both lithologic and faunal basis since the first geological surveys were conducted (Rogers, 1858). Subsequent work continued to refine the earlier stratigraphic assessments (**see figure 3)**, and work by Collie (1903), Ulrich (1911), Butts, (1918, 1931), Field (1919), Kay (1944), and Thompson (1961, 1963) investigated the lithologies and faunas of these limestones and shales resulting in a hierarchy of stratigraphic names. Refinements included separation of the uppermost part of the Trenton as a shale-dominated unit that was eventually referred to as the Antes Shale (Kay, 1944). Early work initially segregated the massive limestone unit at the base as upper Black River, leaving two primary formations in Pennsylvania. These were referred to as the Salona and Coburn Formations by Field (1919). Nonetheless, Kay (1944) recognized the massive limestones (his Nealmont Formation and the "Black River Limestone" of early workers) immediately below the Salona Formation on a faunal and lithologic basis, and considered the Nealmont to be equivalent to lowermost Trenton rocks of New York and Ontario. Thus today in the Valley and Ridge, the Trenton in Pennsylvania is generally considered to be composed of three formations: the Nealmont, Salona, and Coburn. Some recent workers suggest that the Nealmont should be split with the Centre Hall member of the Nealmont becoming part of the Black River Group and the Rodman member of the Nealmont becoming part of the Trenton Group (see Laughrey et al., 2003).

In contrast to the underlying Black River Group, where carbonates demonstrate repetitive successions of at least three different styles of depositional cyclicity produced in shallow-water environments, the sediments deposited in the Trenton Group are substantially less variable. As reported by Gardiner-Kuserk (1988), the facies in the Trenton in the northwestern Ridge and Valley demonstrate a single style of depositional cyclicity characterized by a fining upward pattern characteristic of distal storm deposits and/or turbidity style currents. These are interbedded with calcareous and/or organic-rich shales that were deposited under oxygen-stressed conditions between event deposits. Given this depositional pattern, many former workers including Gardiner-Kuserk (1988) considered Trenton carbonate sediments to have been produced outside of the local basin and transported into deep-water (deep shelf-slope?) conditions from multiple source areas. Given this change, as elsewhere, the Trenton Group of Pennsylvania was deposited under the influences of major tectonic activation in the Vermontian Phase of the Taconic Orogeny.

Trenton Group Contacts

The lower contact of the Trenton Group with the underlying Black River Group has been discussed previously. The upper contact of the Trenton in central Pennsylvania (as in New York), is placed at the uppermost limit of carbonate-dominated strata below the base of the shale-dominated interval of the Antes Formation. Paleoecology, stratigraphy and chronology of the contact have been the subject of several recent investigations (Beares et al., 2002; Lehman et al., 2002). In the western Ridge and Valley, these former authors identified the age of the interval in the base of the Antes Shale as Edenian on the recognition of *C. spiniferus* zone graptolites (**figure 14**). The interval of the contact contains a transitional series of strata ranging from a zone of calcareous shales to argillaceous limestones before a dramatic and sharp change

Figure 14: Photographs of the Trenton Group – Antes Shale contact at Reedsville, Pennsylvania road-cut (PA 322). The top surface of the Coburn is solution pitted by modern weathering of the pyrite-rich black Antes Shale and unconformable with evidence for synsedimentary corrosion and brecciation of the upper Coburn Limestone contact just after deposition. The basal Antes Shale contains an interval of ~0.5 meters of calcareous shales overlain by a 1.5 - 2 meter thick argillaceous limestone before changing sharply into black *C. spiniferus*-bearing shales as suggested by Lehman et al. (2002). The interval also contains K-bentonites that the former authors correlated into New York.

into black, papery shales of the Antes (Lehman et al., 2002). These authors suggested this transitional interval was equivalent to the upper Trenton including the Steuben Formation. This substantiates earlier conodont biostratigraphic studies, which suggested that the Antes Shale in this region was Edenian in age (Sweet & Bergström, 1976). As indicated by Ryder and colleagues (1998) the bases of these shales and their lateral equivalents across the GACB have the highest total organic carbon content of the Upper Ordovician. These high carbon values were produced as the result of a significant oceanic event at the end of the GACB during deposition of the basal Antes interval.

Distribution

Although early studies identified Trenton equivalents in Pennsylvania, the work of Rosenkrans (1934) helped to identify a number of new exposures and through correlation of at least 12 K-bentonite occurrences, he demonstrated the relative age relationships of most units in the BRT interval. Thompson (1963) specifically investigated the Salona and Coburn Formations in detail and correlated sections within the western and central Ridge and Valley region (Nittany Valley to Kishacoquillas Valley). His work recognized significant progradational and retrogradational patterns in the development of facies within the western half of the Ridge and Valley and helped define units within the Salona and Coburn Formations that extended north-south along the western valleys. Within this region, Thompson (1963) identified the New Enterprise and Roaring Spring members of the Salona Formation and the Milesburg and Coleville members of the Coburn Formation. His work correlated these units throughout most of the Ridge and Valley, although he noted the transition of these units to the southeast into the Martinsburg Formation.

In the central and eastern Valley and Ridge especially in the westernmost Cumberland Valley, the limestones of the Trenton Group grade into interbedded shales and siltstones of the

Martinsburg Formation. In this region the Martinsburg has not been differentiated lithologically into units; however, some biostratigraphic zonation has been applied using graptolites (Stephens et al., 1982). This work resulted in recognition of graptolites from the *C. americanus* zone through the *C. spiniferus* zone within which a darker black shale unit is recognized – possibly equivalent to the Antes Shale of the western valleys. Mapping in the Path Valley region by Hoskins (1976) delineated the strata between the Kishacoquillas Valley and the Cumberland Valley as "Reedsville" and did not further differentiate this unit. This region, however, demonstrates substantial inter-fingering between Martinsburg siliciclastic-dominated facies and carbonate-dominated facies of the Salona and Coburn Formations. Recent field reconnaissance in the Willow Hill area demonstrates the continuity of the K-bentonite succession into interbedded limestones, siltstones, and shales of the "Martinsburg" suggesting that with further field work, the members of the Salona and Coburn may be extended.

In the subsurface, Wagner (1966) recognized the Nealmont-Salona-Coburn interval in the subsurface throughout western Pennsylvania and in northern West Virginia. In the region of the Olin Basin (Rome Trough), the Nealmont and Salona interval is considered to be substantially thicker than in the outcrop belt. In northern Erie County, northwestern PA, the Antes Shale is recorded to rest on Salona-type facies, and then on Rodman-style facies to the southwest in northernmost West Virginia. Wagner suggested that the transition to siliciclastic deposition in this region was due to the northeastward progradation of a tongue of siliciclastic sediment derived from the Ozarks. It is clear from his description and relation to Freeman's (1953) shale filled trough of southern Illinois and Indiana and northern Kentucky, that he felt the siliciclastics were entrained within what is today referred to as the Sebree Trough and delivered to the northeast into western Pennsylvania during deposition of the latest Salona and Coburn. To the north into western New York and Ontario, Wagner (1966) correlated the Salona and Coburn with the Verulam and Cobourg (Lindsay) Formations of Ontario, and with the middle to upper Trenton of central New York.

Nealmont Formation

The Nealmont Formation was used by Kay (1943) to contain the interval originally referred to as upper Black River by Butts (1918). The term Nealmont was thus relegated to the strata between the Lowville-like facies (Benner Formation) and the Trenton-like facies (Salona-Coburn Formations). Kay (1944) conceived of the unit as containing three distinct members. The lowest was the relatively coarse-grained Oak Hall Member (now considered to be part of the upper Linden Hall Formation as discussed previously), the middle argillaceous fine-grained calcisiltite member (the Centre Hall Member), and the upper or crinoid-brachiopod calcarenite interval (Rodman Member; **figure 15**). Overall the Nealmont represents the last massive limestone-dominated formation in the Upper Ordovician and represents a transgressive succession overall from shallow to deep-shelf conditions. Rones (1969) removed the Oak Hall Member as it was only locally developed and considered a facies equivalent of the Linden Hall. The remaining two units, the Centre Hall and Rodman Members are fairly widespread throughout much of the Ridge and Valley region even in West Virginia where they are traced into the Dolly Ridge (Perry, 1972, Ryder, 1992). Throughout this region the Nealmont is used for the interval of argillaceous micritic and bioclastic limestones sitting above the relatively pure limestones of the Linden Hall Formation. Moreover throughout this region the unit averages a fairly constant thickness of about 22 meters although it is somewhat thicker to the north of the Nittany Valley and into the Nippenose Valley (Rones, 1969). To the south at Roaring Spring and near Bedford at the Ashcom Quarry, the Nealmont is also slightly thicker with an average of

Figure 15: Lithologies and contacts of the upper Linden Hall through Nealmont Formation. Lower left shows the Valentine-Oak Hall contact at Pleasant Gap where the transition from massive structureless fenestral micrites are overlain by faintly cross-bedded fine grainstones of the Oak Hall. Center left shows facies of the Centre Hall Member of the Nealmont Formation, and upper left and lower right shows the sharp and irregular contact between the Centre Hall and the overlying brachiopod-crinoid grainstones of the Rodman Limestone (at Union Furnace). The center right image shows the typical wavy nodular bedding in the upper Rodman and the upper right image shows the contact interval between the Rodman and the overlying Salona Formation at the Rt. 322 cut at State College.

about 24 meters.

In the Cumberland Valley region, the Nealmont was correlated with the Mercersburg Limestone by Kay (1944). However, the recognition, of five K-bentonites within the

Mercersburg and a faunal association similar to the Linden Hall suggested the Nealmont was younger than the Mercersburg (Craig, 1949). Although the term is not currently recognized, the interval immediately above the Mercersburg Limestone was referred to as the Greencastle Limestone and considered originally by Kay to be representative of the basal Salona. The Greencastle beds were reported as an interval of dark gray to black shales grading upward into sandstones and argillaceous limestones which were finally capped by a coarsely crystalline limestone bed (Bassler, 1919). The introduction of siliciclastic sands and silts reflects the proximity to the Sevier and Taconic basinal areas. Although detailed stratigraphic correlations are yet to be established, biostratigraphic assessments of these early workers show the interval to contain *C. americanus*-zone graptolites, which the underlying Mercersburg does not. Further south, the Greencastle becomes dominated by sandstones referred to as the Oranda Sandstone (Cooper & Cooper, 1946). In Virginia, Harris and colleagues (1994) identified the lower Oranda as belonging to the late *P. undatus* Zone suggesting a *P. undatus* to earliest *P. tenuis* age for the Oranda which is in agreement with the age for the Nealmont to basal Salona.

Centre Hall Member

The Centre Hall Limestone was named by Field (1919) for outcrops in the Centre County region of the Nittany Valley. The limestone was considered to be a zone of impure limestones averaging about 5 meters-thick in the type region and was positioned immediately above the Valentine Limestone and below the Rodman. Kay (1944) considered the Centre Hall to have a distinct massive and coarser lower portion that he separated off at the position of the N2 K-bentonite (figure 16). He referred to this lower unit as the Oak Hall (**figure 16**). Kay relegated the upper portion to the Centre Hall with its base established at the level of the N2 K-bentonite. The Centre Hall extends upward through an interval of thinner-bedded, argillaceous calcilutite

Figure 16: Exposures of the Oak Hall – Centre Hall Member contact and the N2 K-bentonite in the north pit or "White rock" quarry at Pleasant Gap, Pennsylvania. The Oak Hall (below) was separated from the Nealmont by Rones (1969) and considered a lateral equivalent of the Linden Hall. Nonetheless, in this locality, the Oak Hall sits in a position immediately above the Valentine Limestone and below the Centre Hall Limestone. Facies at the top of the Oak Hall resemble the overlying Centre Hall although they are somewhat coarser grained and have a more massive character.

limestones to the base of the thicker (15 to 30 cm) crinoidal calcarenites of the Rodman Member.

Rones (1969) included about 13 meters in the Centre Hall, but indicated it was difficult to draw the upper contact at a specific horizon. At Union Furnace, as shown in figure 7, the Centre Hall Member is approximately 10.34 meters thick, and at Roaring Spring, the Centre Hall Member is about 12.2 meters where it sits above 5.5 meters of Oak Hall lithology.

In the Centre County region, the Centre Hall is composed of two main lithologies. The first is a medium gray (weathering light gray) thin-bedded calcilutite with scattered brachiopods, *Maclurites* and other gastropods, and other fossil debris. This unit typically weathers into 10 to 20 cm thick platy to wavy-nodular beds with argillaceous seams that can be somewhat dolomitic (**see figure 15**). The second lithology is a coarser calcisiltite facies with substantially fewer fossils, but with some evidence of burrowing and vertical boring in hardgrounds at the tops of the beds. These intervals are typically thicker (0.30 – 1.0 meter-thick) and typically alternate in successive cycles with the underlying lithofacies. In addition to the basal K-bentonite, the Centre Hall also contains the N3 K-bentonite of Kay (**see figure 13**). This K-bentonite appears about six meters above the N2 K-bentonite and is intermittently recognized in the State College region, but has also been recognized at Reedsville in the Kishacoquillas Valley and also at Orbisonia in the Black Log Valley (Rones, 1969). The N3 bed is also evident at Roaring Spring where it was used as a parting to form the northwestern side wall of the quarry.

As reported by Kay, the fossils in the Centre Hall were similar to those found in the Rockland of Ontario (Selby – Napanee Formations) and thus on this basis Kay assumed the Centre Hall to be Rocklandian in age. The specific faunas he observed included a number of taxa. In general, fewer corals were documented than in underlying units but the Centre Hall contained the rugose coral *Lambeophyllum profundum*. The unit also has a number of brachiopod taxa including *Doleroides pervetus*, *Hesperorthis tricernaria*, *Sowerbyella curdsvillensis*, and at least ten others. Also mentioned are numerous gastropod taxa including

Maclurites logani, *Subulites sp.*, *Liospira,* and *Lophospira*. Most of these taxa are found in the upper Black River to lower Trenton of New York, and many, including *Subulites*, are found in the Carters Limestone in Tennessee and in other mud-dominated facies of this approximate age (Stanley, 1977). At least five different cephalopods are also listed, including the distinct dorso-ventrally flattened actinocerid *Gonioceras kayi*, which is known from the Decorah of Minnesota and the Guttenberg of Iowa (Kay, 1934). Other forms are known from the lower Black River in Pennsylvania as reported earlier. At least eight taxa of trilobites are known, but do not include *Cryptolithus tesselatus* that is found first in the Salona. A number of ostracods are also identified and correlated again with forms in the Decorah to Guttenberg of Iowa and Minnesota. In the eastern Ridge and Valley, the equivalent unit to the Centre Hall has not been specified although it correlates approximately with the "Greencastle" as discussed previously. Some beds in the Greencastle yield particularly abundant *Subulites,* as reported by Bassler (1919), so these beds may be the lateral equivalent.

Rodman Member

The uppermost interval of the Nealmont Limestone is referred to as the Rodman Member (**figure 17**). First recognized by Butts (1918) in the Roaring Springs Quarry, the Rodman is a ten meter-thick fossiliferous black crystalline limestone weathering to medium to dark gray. The lower contact of the Rodman was established by Kay (1944) as gradational, but generally is placed at the base of the first massive calcarenite or coquinal bed (see figure 15), and the upper contact was drawn at the position of the Salona 0 K-bentonite, although the lithologic contact is usually just below this K-bentonite. Given these contacts, subsequent investigation by Rones (1969) recognized the Rodman to vary in thickness within the western Ridge and Valley between five and ten meters usually thickening at the expense of the Centre Hall. The unit averages about

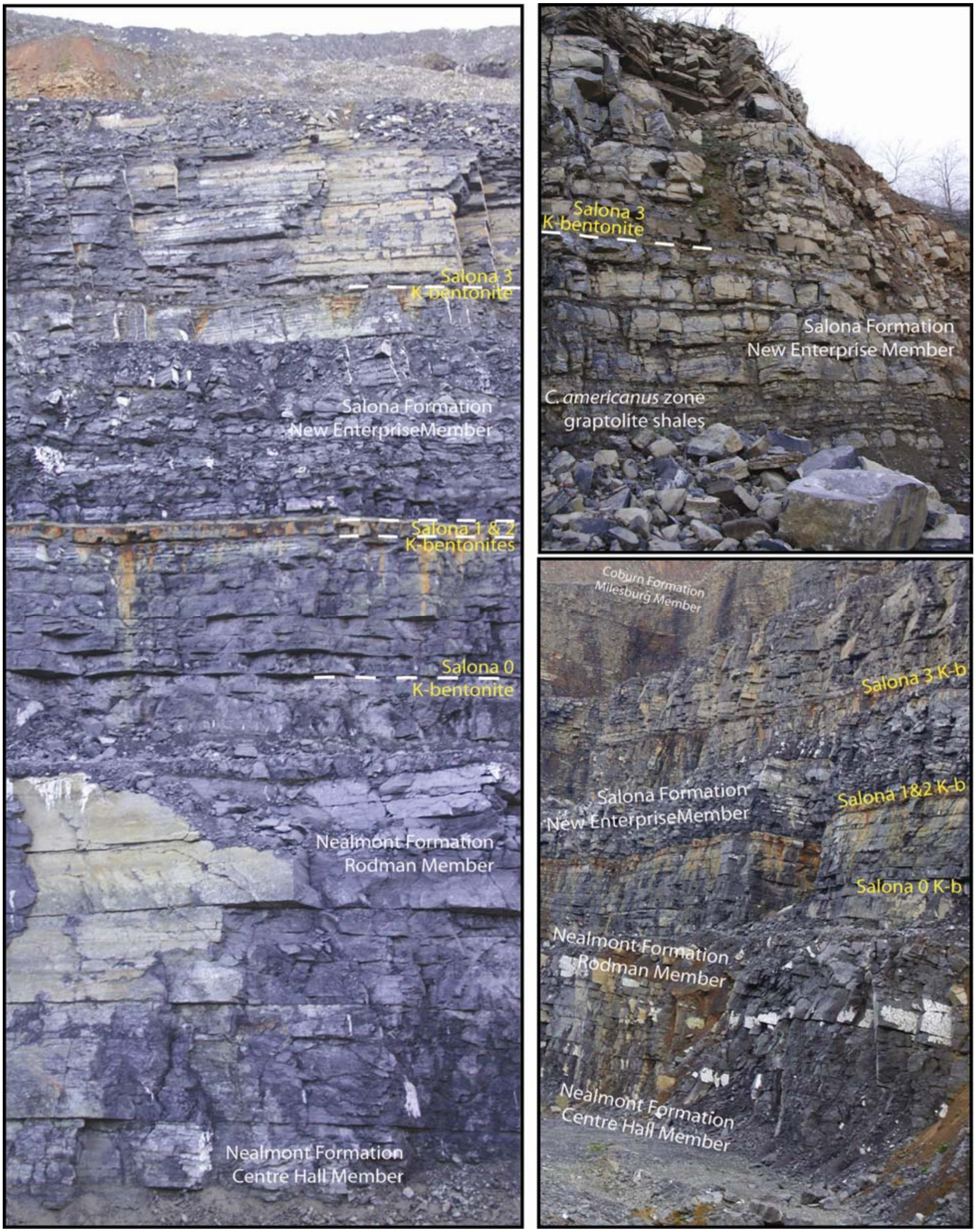


Figure 17: South wall of the north pit in the Glenn O. Hawbaker Quarry at Pleasant Gap showing the interval between the Nealmont and the overlying Salona Formation. In the uppermost part of the quarry, exposures reach the level of the Coburn Formation. Clearly observed, however, is the position of the Salona K-bentonites including the "twins" 1 & 2.

nine meters in the Nittany Valley region. At Union Furnace about 8.2 meters of the Rodman are developed, at Pleasant Gap about 9.75 meters are attributed to the Rodman, and at Oak Hall about 9 meters are exposed. The Rodman thins substantially to the southeast in the section at Reedsville.

Typically the Rodman, like the underlying Centre Hall, is composed of two dominant lithologies. The first is a somewhat argillaceous medium-to-coarse-grained bioclastic calcarenite or packstone. The second is a medium- gray, thin-bedded calcilutite or wackestone facies similar to the underlying Centre Hall (Rones, 1969). Typically, the lower one third of the unit is dominated by interbedded calcilutites and bioclastic grainstones with the upper becoming more dominated by condensed bioclastic calcarenites with easily recognized crinoid fragments, other pelmatozoan pieces, and a large bryozoan component that stands in contrast to the underlying beds. Within the Rodman despite the coarsening of facies, the overall trend is an upward thinning of beds from the medium-bedded planar calcilutites to thin, bioturbated facies that weather to nodular bedding. Organic and siliciclastic components also become somewhat more prevalent with the addition of a noticeable shale fraction and finely disseminated quartz silt. The upper part of the Nealmont is also known for an increase in the abundance of dark grey to bluish black chert nodules that appear to be bedded in some cases. In figure 17, the thicker-bedded lower Rodman is easily seen as is the more rubbly-weathering upper Rodman.

Outside of the Valley and Ridge, the Rodman Limestone was correlated by Kay (1944) with the upper Mercersburg Limestone. Nonetheless, based on a reevaluation of data presented by both Bassler (1919) and Craig (1949), the Rodman likely correlates with an interval in the base of the Martinsburg that has been variously attributed to the Greencastle Limestone (in the Cumberland Valley) or the Myerstown Limestone (in the Lebanon Valley). In southern

Pennsylvania, Bassler described an 11 meter-thick, sub-granular to granular, nodular-bedded limestone and shale unit with noticeable crinoid elements, and some receptaculitids that was immediately overlain by a much finer-grained carbonate interval with the distinct gastropod *Sinuites* and the first occurrence of *Cryptolithus tesselatus*. Given the appearance of the crinoidal wackestones and packstones overlain by fine-grained facies containing *Cryptolithus tesselatus*, it is believed that this interval represents the base of the overlying Salona Formation in this region. Further south in Virginia this interval was previously referred to as the upper Oranda. To the northeast, Gray (1952) named the Myerstown Limestone in the Lebanon Valley for an interval of dark-blue gray to black, organic-rich limestones with interbedded calcarenites that sat immediately below the Hershey Conglomerate. The upper portion of the Myerstown contains at least four K-bentonites, which again likely correlate with those in the Salona. Thickness of this specific unit is not established; however, the total Myerstown is approximately 60 meters thick.

In the subsurface of western Pennsylvania, the Rodman calcarenites were reported by Wagner (1966) to thicken to a maximum of 30 meters or more, but considered it possible that the Rodman calcarenites moved up section and thus were time equivalent of younger units in the Valley and Ridge. Specifically, he correlated the closely-spaced K-bentonites of the New Enterprise member of the Nittany Valley region with a cluster of K-bentonites at the top of his Rodman equivalent in northwestern Pennsylvania. Within this upper unit, Wagner indicated that the Rodman showed significant evidence for shallowing-upward of facies which is contrary to the pattern of deepening observed in outcrops in the Ridge and Valley. Thus it is considered likely that at least part of the "Rodman" of western Pennsylvania is younger than the Rodman of central Pennsylvania. Moreover, based on his correlations of this interval into New York (into the Kesselring Well, Chemung County) the equivalent unit was considered to be Kirkfield. Thus based on correlation of K-bentonites in wireline logs and cores, it appears that this "younger Rodman" of northwestern Pennsylvania is not time equivalent to the type Rodman, but is the same age as the **upper** New Enterprise Member of the Salona and is equivalent to at least part of the Kirkfield Limestone (see below).

Biostratigraphically the Rodman Member, as compared to the underlying Centre Hall Member is known to contain only one coral taxon (*Lichenaria coboconkensis*). Instead, the Rodman is dominated by a bryozoan fauna including multiple species of the fenestrate bryozoan *Chasmatopora (Subretopora) reticulata* and *C. sublaxa* (Kay, 1944). Other members of this genus are also known from Chazyan to earliest Black River in siliciclastic-influenced facies in Tennessee (Upper Murfreesboro, Pierce, and Lebanon Limestones) and Virginia (Ward Cove and Wardell Limestones) and from Rocklandian strata of eastern New York (Glens Falls Limestone). Thus the occurrence of the *Chasmatopora* in the Rodman likely represents a recurrent fauna that is relatively coeval with the occurrences in the New York sections and coincident with the onset of siliciclastic sedimentation in earliest "Trenton time." Brachiopods are also common, with up to seven different forms recognized including *Sowerbyella curdsvillensis* as noted previously. Interestingly, *Hesperorthis tricernaria* has not been recognized nor has *Dalmanella* sp., both of which are common in equivalent strata elsewhere. The Rodman has also produced relatively few trilobites and ostracods compared to the underlying member of the Nealmont. Although some taxa were omitted from the Rodman, based on the echinoderms and correlation with the other areas, Kay originally believed the Rodman to be equivalent to the Kirkfield. Nonetheless, as shown here and by correlations of other workers, the Rodman is likely older than the type-Kirkfieldian and lithologically and faunally more similar to portions of the Rockland of New York and Ontario.

Salona Formation

The Salona Formation was named for exposures in the northern Nittany Valley by Field (1919) for the dark, argillaceous, fine-grained limestones immediately overlying the Rodman Member. In the type region, Field identified nearly 75 meters of interbedded fine-grained limestones and shales below the coarser but shalier interbedded limestones of the Coburn Formation. Following the precedent established by Whitcomb (1932), Kay (1944) attributed a somewhat smaller thickness to the unit in the type section (~53 meters). Kay and Whitcomb favored a base for the Salona at the level of the first calcilutite above the nodular coarse-grained Rodman Limestone. In Salona, this bed occurred approximately 60 centimeters below the Salona 0 K-bentonite, although the base of the Salona 0 K-bentonite has been commonly used as the base. Kay (1944) considered this contact to be relatively conformable, although he also noted that a dramatic syndepositional angular unconformity was present in the quarry near Antes Gap in the Nippenose Valley. In this location, the Nealmont interval was tilted by faulting and is overlain with angular discordance by the basal beds of the Salona Formation. Although not at the base, there is evidence for deformation and synsedimentary faulting in the lower New Enterprise on PA Route 322 at Oak Hall as well (see below). This strongly suggests that significant tectonic activity was initiated coincident with onset of deposition of the Salona in this region.

The top of the Salona was established at the top of the relatively thin-bedded limestones of the Salona where it graded into coarse-textured, coquinal limestones of the Coburn Formation.

As defined the Salona contains a number of persistent K-bentonites (Salona 0, 1, 2, 3, 4, and 5 after Whitcomb, 1932, **see figure 17**) although Kay (1944) recognized potentially upwards of eight K-bentonites in the interval. After recognition and correlation of a number of K-bentonites within these units by Whitcomb (1932) and Rosenkrans, (1934), Thompson (1961, 1963) was able to further correlate and differentiate both the Salona and Coburn into two members each. The Salona was subdivided into a basal New Enterprise Member and an upper Roaring Springs Member which are recognizable over most of the Ridge and Valley region except for in the eastern Cumberland Valley region (Thompson, 1963).

As reported by Thompson, in the type region at Roaring Spring (the New Enterprise Quarry), the lower unit is dominated by structureless calcilutites and the upper is composed of interbedded laminated and cross-laminated calcilutites and fine-grained calcarenites (**figure 18**). Faill and colleagues (1989) report the New Enterprise Member to consist of relatively unfossiliferous, interbedded, dark-gray to grayish-black calcisiltites and calcareous shales. They distinguish the Roaring Spring member by the presence of ripples and cross-bedding in an interval of interbedded calcisiltites and calcarenites. Historically, only one K-bentonite bed was documented in the Roaring Spring as opposed to five in the underlying New Enterprise; however, Berkheiser and Lollis (1986) suggest there might be as many as six K-bentonites in the Roaring Spring (see figure 7). The former authors measured approximately 55 meters in the New Enterprise quarry at Roaring Spring, while about 56 meters are present at Union Furnace (60 meters was reported by Laughrey et al., 2003). As reported by Thompson (1963) nearly 54 meters of Salona are present in the outcrop on PA Route322 between State College and Oak Hall (**figure 18**). Unfortunately, the upper member of the formation is not fully exposed at the type section at Salona. This fact might in part explain the discrepancy between the 75 meter-

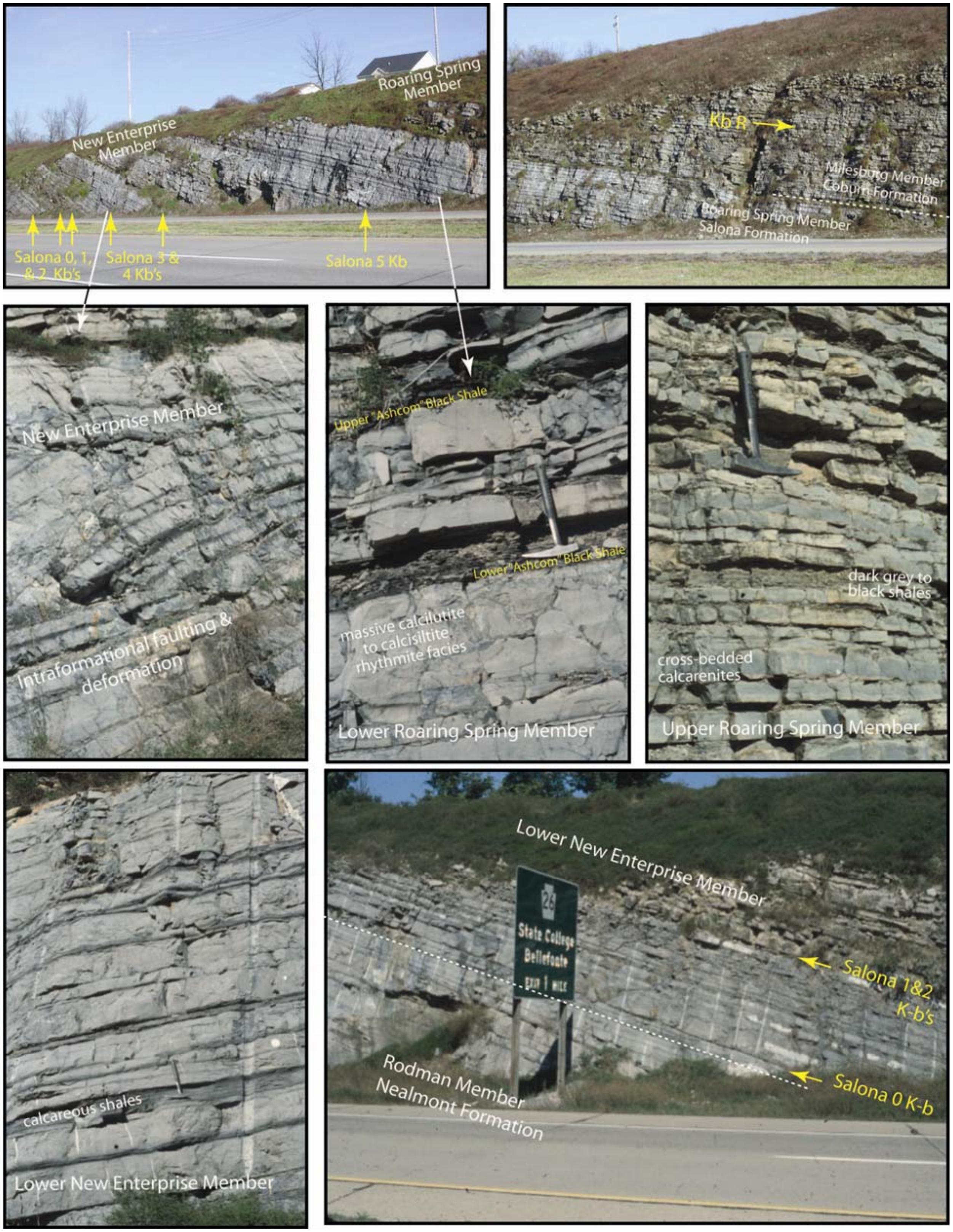


Figure 18: Outcrop images for the Salona Formation as exposed along PA Route 322 1 mile east of the exit for Rt. 26 to State College in the vicinity of Oak Hall. Images show the typical facies and features of the Salona Formation.

thick Salona as measured by Field versus the 55 meters of other workers – it is likely that Field included some of the lower Coburn in his early measurements.

Outside of the Ridge and Valley area of Pennsylvania, the Salona has long been correlated with the lower Martinsburg both in Virginia and West Virginia. Ryder and colleagues (1992) correlated the Upper Ordovician succession in the subsurface between northern Virginia and Ohio and recognized the Salona Formation and its equivalents to the west. In the subsurface of eastern Ohio, the Salona was established to be Rocklandian in age based on conodonts and the correlation of K-bentonites. This age estimation is generally in agreement with the age assessments of Sweet and Bergstrom (1976). To the east of the Cumberland Valley, the Salona has not been explicitly correlated although a number of K-bentonites have been recognized in the top of the shaly limestones of the Myerstown Formation and at least a couple of bentonitic horizons are noted in the deformed conglomerates in the base of the overlying Hershey Limestone in the Lebanon Valley just east of the Susquehanna River. Above the distinctive Hershey conglomerates, the limestones disappear rather suddenly as the unit grades into the faulted shales and siltstones of the Dauphin Member of the Martinsburg Formation (Willard, 1943). South of the Susquehanna River in the Cumberland Valley, due to faulting, extreme folding, and deformation few exposures show a complete unfaulted interval above the Greencastle (Craig, 1949). Nonetheless, outcrops of the lower Dauphin Member of the Martinsburg are recognized on the basis of graptolites which indicate these interbedded siltstones and shales are uppermost *D. bicornis* to lower *C. americanus* zone and are thus representative of latest Turinian to Rocklandian.

New Enterprise Member

As the lower member of the Salona Formation, the New Enterprise was defined by Thompson (1963) near Roaring Spring as composed of a nearly structureless interval of calcilutites to calcisiltites that on fresh surfaces are nearly black owing to the substantial organic matter contained in the limestones (**figure 19**). As shown in figure 18, the Salona weathers to a

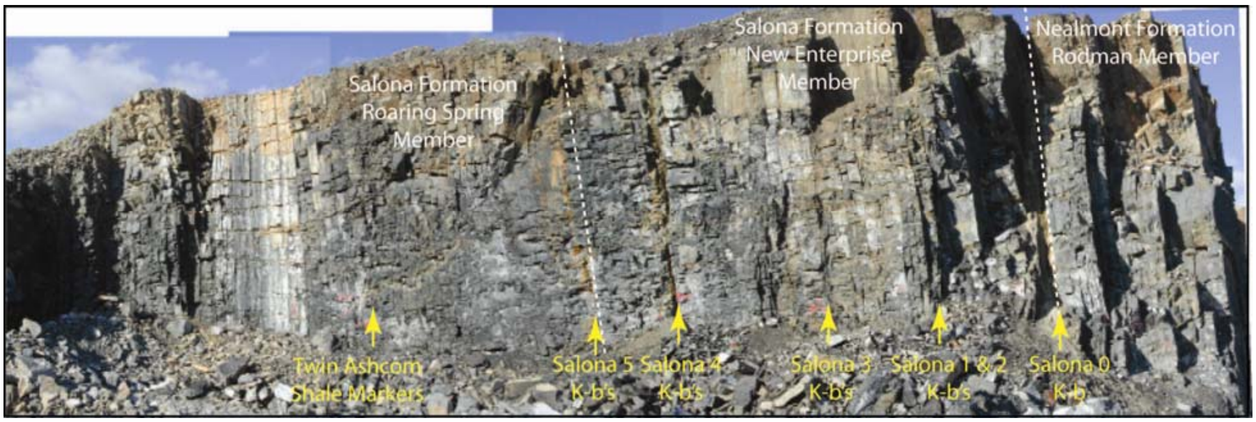


Figure 19: Exposure along the northwestern wing of the New Enterprise Quarry at Rodman, near Roaring Spring, Pennsylvania. Shown in this view is most of the Salona Formation – the top of which is exposed to the left of the image. The positions of the Salona K-bentonites are shown as are two relatively prominent shale beds in the lower part of the Roaring Spring Member that are informally referred to as the Ashcom shale markers herein.

light to medium grey. Beds are characteristically 15 to 30 centimeters thick, but some are substantially thicker. Like the underlying Nealmont, the New Enterprise maintains a remarkably consistent thickness over much of the western Ridge and Valley region where it has been measured and has been referred to as a fairly uniform "blanket-like" deposit of fine-grained structureless calcilutites. In the type locality, the New Enterprise is just over 27 meters, and decreases to just under 26 meters at Union Furnace and between 26-27 meters at Oak Hall and Pleasant Gap (see figures 18 and 17 respectively). In the Black Log Valley, the upper contact of the New Enterprise is reported by Thompson (1963) to move up section as the unit interfingers with coarser-grained Roaring Spring-like facies on top of the Adirondack Axis thus in this region just over 48 meters of strata are attributed to the New Enterprise. In Path Valley at Willow Hill just a short distance to the south, the same interval again thins to just under 27 meters suggesting that the Path Valley was located east of the Adirondack Axis and in slightly deeper water. The increased thickness on the arch is thought to reflect slightly higher carbonate production rates afforded in the slightly shallower water.

Lithologically, as reported by Faill and colleagues (1989), the New Enterprise is composed of nearly black, calcisiltites and interbedded calcareous shales and is generally distinguished from the overlying Roaring Spring Member by the cross-bedded calcarenites often capped with rippled surfaces of the latter. Slupik (1999) also noted after Thompson (1963) that most calcisiltite beds in the New Enterprise are often welded into medium –to-thick beds that occasionally contain thin skeletal lags, and argillaceous mudstone partings. On etched surfaces, it is possible to discern graded bedding with scoured bases. As reported by Gardner-Kuserk (1988), beds in the New Enterprise intermittently display a thin bioclastic lag containing brachiopod, crinoid, bryozoan, and trilobite fragments that grade upward into calcilutites which are finally capped by the shales. Although structureless in appearance, the calcilutites are usually finely laminated and occasionally display some evidence for hummocky cross-stratification indicating the influence of storm deposition. The repetition of this pattern yields the typical “rhythmite” pattern.

In the outcrop regions, the New Enterprise beds become thicker-bedded, and somewhat less fossiliferous to the south toward the region of the Ashcom Quarry at Bedford and to the east into the Black Log Valley (Thompson, 1963). Typically in the north, calcilutites dominate, forming just over 90 % of the member, and are typically interbedded with thin shaly seams. To the south and east the beds become somewhat more amalgamated and demonstrate evidence for multiple reworking events. Amalgamated lutite packages are, in turn, separated by thicker shale units indicating a slightly higher-rate of background siliciclastic sedimentation than farther north.

The presence of thicker amalgamated and basally scoured beds, suggesting perhaps a slightly more energetic regime and/or slightly shallower water conditions than existed in the north – these observations suggest that the Adirondack Axis still played a role in influencing deposition albeit in a much deeper water setting than during deposition of the upper Black River Group.

In the subsurface to the west of the Ridge and Valley, Wagner (1966) correlated the facies of the Salona into a substantially thinner interval on the Kane Arch in northern West Virginia where the New Enterprise grades into somewhat more coarse-grained calcarenites that were classified by Wagner as Rodman lithologies. In this region only about nine meters of Salona are recognized. Although very few calcarenites are found in the New Enterprise in the Ridge and Valley, Thompson (1963) indicated the development of calcarenite beds above the Salona 0 K-bentonite at Union Furnace. These same beds have been recognized by Slupik (1989), but as these are relatively subordinate to the calcilutites they have received relatively little attention. Nonetheless it appears that these beds thicken and become dominant to the south west into shallower waters in the vicinity of southwestern Pennsylvania and northern West Virginia. In contrast in northwestern Pennsylvania in the vicinity of Erie County, the Salona interval is represented by about 25 meters of typical facies with the upper interval being replaced by substantially more shaly facies described as typical Utica by Wagner. It appears that the "Salona" beds of northwestern Pennsylvania are actually equivalent only to the Upper Salona – Roaring Spring Member. To the south and east of the Ridge and Valley outcrops studied by Thompson (1963), the New Enterprise Member appears to pass laterally into the Jacksonburg Limestone Formation and its equivalents, including the Hershey Conglomerate.

Faunally, the New Enterprise Member has been described as a relatively low-diversity interval. Most forms appear to have been substantially comminuted and show evidence of significant fragmentation and corrosion. Nonetheless, some relatively robust forms are abundant on some bedding planes. As reported by Kay (1944) one of the key taxa in the lower New Enterprise is the pelmatozoan *Echinosphaerites aurantium.* It is found with occasional crinoid columnals. Other forms include the trilobites *Isotelus* sp., *Cryptolithus. tesselatus*, *C. concavus*, *Tetralichas trentonensis* (*Lichas trentonensis*), *Flexicalymene senaria*, *Illaenus sp*., *Echarpes,*sp. and *Homalonotus trentonensis*, the trepostome bryozoans *Prasopora simulatrix*, and *P. orientalis,* and a number of brachiopods including *Leptelloidea pisum*, *Rafinesquina*, *Dalmanella rogata*, *Onniella* sp., *Sowerbyella curdsvillensis*, and *Leptaena* sp. (Kay, 1944; Thompson, 1963). Kay (1944) also recognized the cephalopods *Sinuites cancellatus*, *Geisonoceras tenuistriatum*, and *Trocholites ammonius* – the latter of which are frequent to abundant in one bed just below the Salona 2 K-bentonite. Graptolites are also occasionally recovered from the New Enterprise just above the Salona 2 K-bentonite and the Salona 4 K-bentonite and include *Nemagraptus mohawkensis, Rectograptus amplexicaulis,* and *Normalograptus brevis-strictus (Climacograptus strictus)* respectively.

As suggested originally by Kay (1944), this faunal assemblage is typical of the Shorehamian or Sugar River to basal Denley of New York State which was considered by him to be time equivalent to this unit. Nonetheless, the faunas are also permissive of correlations with the early *C. americanus* graptolite zone and Rockland through Kirkfield strata which contain the first occurrence of this deep-water assemblage with a second and somewhat more widespread recurrence in overlying strata. The abundance of *Cryptolithus* in the very base of the New Enterprise followed by the outage of the trilobite upward in the unit, combined with Wagner's (1966) distinct shallowing or progradation of facies in the subsurface of western Pennsylvania at the level of the upper Salona K-bentonites suggests that the New Enterprise Member may represent the deepest phase (fault enhanced?) of the first Trenton depositional highstand followed by subsequent shallowing into the upper New Enterprise. This pattern is thus coincident with the trends observed in the Rocklandian to Kirkfieldian interval of New York and Ontario.

Roaring Spring Member

The upper member of the Salona Formation was termed the Roaring Spring Member by Thompson (1963) for exposures in the quarry near Roaring Spring in the western Ridge and Valley. The unit was defined on the basis of a significantly greater proportion of laminated-to-cross-laminated, fine-grained calcarenites that are interbedded with calcilutites and shales in the upper part of the Salona. The calcarenites or fine-grained grainstones and packstones constitute approximately twenty percent of the unit while thin calcilutite facies comprise another sixty percent. The dark calcareous shales make up the remaining component of the member. Thompson recognized a total of approximately 23 meters in the type locality where the Roaring Spring Member underlies the Milesburg Member of the Coburn Formation. This is in contrast to the 18 meters recognized by Faill and colleagues (1989). Overall the Roaring Spring when compared to the underlying New Enterprise is marked by a significant decrease in bed thickness, a significant increase in argillaceous components, and a change to laminated, quartz-rich lime mudstones as the dominant lithology.

Unlike the underlying New Enterprise, the Roaring Spring Member appears to be somewhat less rigidly defined and the unit has a considerably variable thickness – especially with respect to its upper contact. The base of the Roaring Spring member is drawn at the level of the first abundant cross-bedded calcarenites. It is clear from Thompson (1963) that the base of the calcarenites lay about at the level of the Salona 5 K-bentonite. In his correlated cross-sections, he often placed the base just below the Salona 5 K-bentonite and to the southeast onto the Adirondack Axis, he put the base down at a level a couple of meters below the Salona 5 K-bentonite. Further east in the Path Valley, the base of the calcarenites was placed about one meter above the Salona K-bentonite. Despite the somewhat variable level, the occurrence of the Salona 5 K-bentonite helps constrain the lower contact of the Roaring Spring Member and is relatively easy to correlate on this basis following the work by Whitcomb (1932).

The upper contact of the Roaring Spring Member is somewhat more difficult to place and is generally variable in its position relative to specific time planes. Field (1919) originally recognized the contact between the Salona and the Coburn at the appearance of highly fossiliferous wackestones, packstones, and brachiopod-rich rudstones interbedded with black shales and calcilutites. These fossiliferous beds stand in sharp contrast to the underlying Roaring Spring Member that is typically quite unfossiliferous. Thompson (1963) thus drew the top contact of the Roaring Spring at the base of the first *Sowerbyella*-bearing limestone pavements.

In the Bellefonte region the top of the Roaring Spring Member was established at a level approximately 25 meters below the position of a prominent yellow to white-weathering K-bentonite referred to as the R K-bentonite by Thompson (1963). This gives an overall thickness to the Roaring Spring Member of approximately 30 meters. Thompson recognized somewhat less at Oak Hall (approximately 24 meters). To the north, however, at Coburn, the first *Sowerbyella* wackestone and packstone pavements, and thus the base of the Coburn Formation, do not occur until a few meters **above** the R K-bentonite. Thus at Coburn, the Roaring Spring is considered to be somewhat thicker and time transgressive with a thickness of about 53 meters. Similarly southward into the Kishacoquillas Valley at Reedsville, the contact between the

Roaring Spring and the *Sowerbyella*-bearing Coburn Formation climbs in the section to a level 18 meters above the R K-bentonite. Thus with a nearly level and synchronous base, and a time-transgressive upper contact, the Roaring Spring Member is substantially thicker with values between 50 to 76 meters in the region between Reedsville and the Shade Gap in the Path Valley. Throughout this region however, the Roaring Spring maintains a fairly uniform lithologic appearance with the exception that the unit becomes more argillaceous in the central and eastern Ridge and Valley where it finally transitions into Martinsburg lithologies in the Cumberland Valley. In this general direction beds of the Roaring Spring member also become somewhat thicker with more calcilutite interbeds compared to the thinner calcarenite cycles to the northwest.

In the subsurface, to the southwest of the Ridge and Valley, Wagner recognized an equivalent facies of the Roaring Spring near the Pennsylvania – West Virginia border but notes that the 30 meter-thick unit is finer-grained and describes the interval as a calcisiltite-dominated unit as opposed to calcarenitic. To the northwest Wagner suggests that the contact between the Coburn and the Roaring Spring continues to drop in the section as was the pattern in the Ridge and Valley. Thus in western Pennsylvania the fossiliferous facies typical of the Coburn appear at lower levels and facies grade into those typical of the Verulam outcrops in southern Ontario. Wagner considered it better to apply the Ontario nomenclature to this interval and suggested the upper Salona facies in this region correlated into the Verulam Formation and felt that the Coburn correlated into the Lindsay in this region. In westernmost Pennsylvania, the Roaring Spring equivalent became dominated by shales and was attributed to a “Utica-type” facies. To the north into New York, Wagner felt that the Salona (at least the upper Salona) could be correlated with the Shoreham (Sugar River) and Denmark (Denley) of New York State.

As discussed, the Roaring Spring Member is characterized by a relatively low diversity, low abundance fauna. Thompson (1963) recognized occasional specimens of *C. tesselatus* at the very base of the member and just above the R K-bentonite at Reedsville. He also denoted the occurrence of the brachiopod *Parastrophina hemiplicata* on some bedding planes near the top of the formation as well as the trilobite *H. trentonensis* at the base of the formation, usually just below the Salona 5 K-bentonite. Other forms include a few specimens of typical *Rafinesquina*, and at least one new species is also noted from near the top of the member. In addition *Leptaena*, *Sowerbyella*, *Dalmanella*, and *Prasopora* are also found occasionally in storm-influenced assemblages with these forms becoming more prevalent upward toward the contact with the Coburn. Kay (1944) does not report any additional taxa from this interval. Overall the fauna is taxonomically consistent with a Trenton fauna. In Ontario and New York, although *Parastrophina hemiplicata* is known intermittently from an extended interval from the Kirkfield upward into the lower Lindsay, there are a few zones where it becomes especially abundant including in the middle Verulam and again in the Lindsay (Liberty, 1969). The incursion of *P. hemiplicata* near the top of the Roaring Spring suggests a change in facies that may indeed coincide with one of these zones of increased abundance the Verulam of Ontario. In turn, the latter may coincide with the latest Poland to earliest Russia members of the Denley Formation in New York.

Coburn Formation

The uppermost unit for discussion in this study is the Coburn Formation. The Coburn Formation was recognized by Field (1919) in the Nittany Valley near Bellefonte as an interval of upwards of 125 meters of highly fossiliferous, crystalline limestones and black shales. Kay (1944) considered the Coburn to be a coquinal limestone unit immediately above the Salona and below the black Antes Shale. Faill and others (1989) report a maximum thickness for the Coburn in the Union Furnace region to be about 95 meters. Unfortunately, most natural exposures of the Coburn are highly weathered, especially where in contact with the Antes, and/or are folded, faulted, and deformed making stratigraphic study of the unit very difficult, except in recent road cut exposures including those at Union Furnace on PA Route 453 in the western Ridge and Valley and at Reedsville along PA Route 322 in the Kishacoquillas Valley. Additional exposures are also afforded in a number of limestone quarries. However, the Coburn itself is often not quarried or exposed to its full extent due to the high siliciclastic content and the appearance of occasional pyrite-rich layers. These facts presumably account for the relatively few detailed stratigraphic studies investigating these units. To date, the stratigraphic synthesis of Thompson (1963) provides the only detailed analysis available for the Coburn. In this study, Thompson recognized two main members within the Coburn and referred to these as the Milesburg (lower member) and the Coleville (upper member).

The Milesburg Member is usually recognized on the occurrence of abundant K-bentonites (at least four to five K-bentonites have been recognized including the widely recognized R K-bentonite of Rosenkrans (1932) along with numerous cross-bedded grainstones dominated by *Sowerbyella* and trilobite coquinas. In contrast, the overlying Coleville becomes dominated by *Dalmanella*-and crinoid-rich coquinas within an overall finer matrix – usually calcisiltites to calcilutites. These in turn grade upward into *Cryptolithus*-bearing units that eventually give way into *Triarthrus* bearing black shales in the base of the Antes. Thus, overall, the package shows an upward fining trend with a coincident increase in thickness of shale interbeds.

The stratigraphic boundaries of the internal members are recognized after Thompson (1963), yet very few recent publications apply these terms owing to the complexity of establishing the specific boundaries. Nonetheless as defined, the base of the Coburn is drawn at the level of the first significant *Sowerbyella* packstones and rudstones, generally below the level of the R K-bentonite. Although inconsistently recognized around the region, at least two to three K-bentonites (K-bentonites 18, 19, and 20 of Berkheiser and Lollis, 1986) are found in the immediate vicinity of the first *Sowerbyella* rudstones, and at Union Furnace a massive-bedded medium grainstone bed with numerous large, but disturbed *Prasopora* colonies is considered to be the base of the Coburn by Berkheiser and Lollis (1986).

The upper contact of the Coburn Formation was placed by Kay (1944) in the vicinity of the first lowest thick, fissile dark shale, but he recognized the transition from the Coburn into the Antes as somewhat gradational. Faill and colleagues (1977) suggested that the contact interval extended over a zone of approximately 15 meters, but Fail and colleagues (1989) used the first major thick, homogeneous, black shale with moderate to excellent fissility, immediately above the level of the last fossil-rich beds of the Coburn. Thompson (1963) drew the upper contact just above the level of the last *Dalmanella*-bearing beds and just below the first beds of the Antes bearing *Triarthrus eatoni*. This position was also supported by Swartz (1955) who also recognized the graptolite *Dicranograptus nicholsoni* in the overlying Antes – a graptolite which is reported by Goldman and colleagues (1994) to be representative of the *C. spiniferus* graptolite zone.

Thus, overall the Coburn is composed of thin-to-thick-bedded, dark-gray to black wackestones and packstones with calcisiltite and calcilutite matrix. Some beds also develop into fine-grained grainstones with coquinal rudstones. The only additional distinct associated lithology consists of occasionally large lenses of black chert within the Coburn (Faill et al., 1989). The nodules and lenses range up to 60 cm in length in some levels and mainly occur in the Milesburg Member above the position of the R K-bentonite (Thompson, 1963). As reported by Thompson (1963), given these approximate contacts, in the Centre County to southern Clinton County region where the majority of outcrop sections exist, the Coburn reaches a maximum of about 98 meters in thickness at Bellefonte with the greatest thickness applied to the overlying Coleville Member (-58 meters). To the southwest into Blair County to Roaring Spring, the Coburn is only about 52 meters-thick in total; thus it thins substantially and is nearly equal in thickness to the underlying Salona in this region. Eastward into the Kishacoquillas Valley at Reedsville, the Coburn was measured to be about 18 meters-thick, but this value appears to be in error as Slupik (1999) reported nearly 60 meters of Coburn at the PA Route 322 highway cut although in should be stated that a deformed and faulted interval occurs in the Coleville Member in this locality. Further south into Franklin County in the Path and Cumberland Valleys and in northern Virginia, the Coburn as a limestone unit is essentially unrecognized in the transition off the Adirondack Arch into the Martinsburg Shale in what Thompson referred to as the Champlain Trough –presumably after Kay (1948). Further to the east, previous work on the Martinsburg Formation in the interval below the Reedsville has produced fossiliferous assemblages with communities similar to the fossiliferous *Sowerbyella*-dominated intervals in the Coburn both at Swatara Gap (Lebanon Valley) and in the western part of the Lehigh Valley (Lehman & Pope, 1989; Bretsky et al., 1969). The stratigraphic level suggests the potential for correlation with the Coburn.

In the subsurface Wagner (1966) identified the Coburn to be thickest in the vicinity of Bellefonte and indicate that the unit thinned in all directions suggesting the Bellefonte region was a major depocenter for the unit as was the case for the underlying Salona (Thompson, 1963). In the northwestern part of Pennsylvania, the Coburn Formation was correlated into the Cobourg Formation. Wagner preferred to use the term Lindsay in the subsurface; however, it is likely an equivalent of the upper Verulam to lower Lindsay (Hallowell Member) only or the Rust to Steuben Formations of New York. Farther south in central western Pennsylvania, the Coburn is reported to be absent where Utica Shale facies extend downward and rest on Salona-like facies and in places, even the Salona facies are reported to be missing as in the Davidson and Hockenberry wells in Mercer and Butler Counties, respectively. In these cases, the Utica facies sits directly on Rodman-like grainstones. This pattern suggests that the region immediately south of Lake Erie and west of the Kane Arch may have been influenced by development of the Sebree Trough during earlier Trenton time – a view favored by Wagner who believed the sediments were derived as the northeastward extension of the "Ozark clastic tongue." Hence the succession is remarkably similar to that found in the Sebree Trough of western Ohio.

Faunally, very few studies have separated or distinguished the components of each of the members of the Coburn Formation so discussion here is based on faunal assessments for the entire unit. As reported originally by Field (1919) one of the original components of the Coburn was the return of the lace-collared trilobite, *C. tesselatus* along with the reappearance of *Triplecia* sp. Swartz (1955) further added the appearance of *Sowerbyella sericea*, *Rafinesquina alternata*, *R. deltoidea*, *Dalmanella rogata*, *Ceraurus pleurexanthemus*, *Isotelus* sp., *Flexicalymene senaria*, and *Parastrophina hemiplicata*. In her study of the bryozoan faunas of the Salona and Coburn (Arens, 1988) recognized the facies in the Coburn to be somewhat shallower overall than the underlying Salona as suggested by increased occurrences of scouring and fragmentation of numerous bryozoan colonies including *Prasopora simulatrix* and small twig-style *Eridotrypa* bryozoans. Among her fragmented assemblages, Arens noted the occurrence of other *Prasopora*-like forms, including *Mesotrypa orientalis*, *Cyphotrypa pachymuralis*, *Diplotrypa westoni*, and *Batostoma* cf. *sheldonensis*. As noted by Arens *C. pachymuralis*, *B.* cf. *sheldonensis* and *D. westoni* are characteristic of either the upper Trenton of New York or from the Cincinnati Arch region. *M. orientalis* is only known from the Nashville Dome region and Estonia. Moreover, as shown by Titus (1992) the brachiopod *S. sericea* is constrained to the upper Trenton of New York, including the Rust through Steuben interval (this is a change from his 1986 publication which considered all Trenton sowerbyellids as *S. sericea*). Thus recognition of these forms in Pennsylvania again suggests that the Coburn is coeval with the Rust and Steuben of New York, a correlation which was suggested by Arens (1988).

Milesburg Member

The Milesburg Member of the Coburn Formation is named for outcrop sections in the region immediately north of Bellefonte in the Nittany Valley. Thompson (1963) defined the Milesburg Member on the basis of abundant *Sowerbyella* bioskeletal calcirudites, *Sowerbyella*-trilobite bioskeletal calcirudites, calcarenites, and interbedded calcareous shales. Thompson suggested that the unit formed a wedge-like body that thinned in all directions from Bellefonte as the upper contact of the underlying Salona climbs in the section. In the Kishacoquillas Valley and to the south near Bedford (in the Ashcom Quarry) the Milesburg Member is so substantially thinned that it is not recognized confidently. This description and the correlated cross-sections of Thompson combined suggest the Milesburg Member behaves much like a progradational unit originating in the north and prograding to the south and east. In sections where the Milesburg is well-developed, the lower beds below the R K-bentonite are thin, shaly and dominated by *Sowerbyella* rudstones. These beds are in turn capped by pack-to grainstone beds with a more fragmented *Sowerbyella*-trilobite association, typically above the R K-bentonite. This upper unit is typically thicker-bedded, less shaly, and has a much more uniform thickness than the underlying unit across the region. In most sections, this upper Milesburg sub-unit is between 15 to 20 meters-thick (Thompson, 1963).

Given this lithologic range the total thickness of the Milesburg Member is reported by Thompson (1963) to range between 36 and 50 meters in the western Ridge and Valley from Blair, Centre and Clinton Counties. To the southeast the Milesburg thins substantially in eastern Clinton County near the type Coburn section where Thompson identified only about 20 meters while at Ashcom to the south only 13 meters are recognized. In most cases, it is likely that only the upper Milesburg sub-unit is expressed. In the Kishacoquillas Valley at Reedsville, Thompson did not record any Coburn. Nonetheless the work of Slupik (1999) at Reedsville suggests there is a total of about five to six meters up to the level of her B-11 K-bentonite marker and again field study suggests these beds are similar to the upper sub-unit of the Milesburg. Further south, no Milesburg is reported.

Coleville Member

The Coleville Member of the Coburn Formation is recognized in exposures north of Bellefonte. Thompson (1963) considered the Coleville Member to be dominated by shaly interbedded bioskeletal calcirudites. In contrast to the underlying Milesburg Member, the Coburn is characterized by a greater percentage of crinoid columnals and plates as well as a couple of species of *Dalmanella* brachiopods. These appear again as rhythmites in association with fine-grained calcarenites, calcilutites, and increasingly organic rich calcareous shales. At Bellefonte, Thompson considered the unit as a tri-partite package. The lower three meters is dominated by crinoidal bioskeletal calcirudites, the middle unit (about 8 meters-thick) contains *Sowerbyella* calcirudites similar to the underlying Milesburg, and the upper 40 plus meters are distinguished by interbedded *Dalmanella* and crinoid rich calcirudites.

As shown by Thompson, the Coleville has its maximum thickness in the Centre County region and thins to the south and east where it wedges out and grades laterally into shale-dominated facies of the Antes Shale. A maximum thickness for the unit has shifted northeast of Bellefonte to the region of the type Salona where nearly 58 meters of Coburn are recorded below the first thick, homogeneous Antes Shale bed. At Bellefonte less than 53 meters is developed and elsewhere the Milesburg becomes sharply thinner (only 8 meters was reported for the Roaring Spring section) although upwards of 37 meters are now recognized below the first massive shale. Slupik (1999) recognized just over 40 meters of Coleville at Reedsville. Based on these data –the basal Coleville is the most widespread with uppermost beds becoming substantially shalier and transitioning into Antes-like facies to the south and east. This pattern suggests a retrogradational or back-stepping pattern for the Coleville. In the Reedsville section, beds just above the Coleville produce *Triarthrus* sp. trilobites forms not found in the Coburn itself.

Chapter 5: Stratigraphic Summary of the Ashbyan to Mohawkian Interval of the Cincinnati Arch, Jessamine Dome Region.

ABSTRACT

The late Mohawkian to Cincinnatian interval of the Cincinnati Arch-Jessamine Dome region has been well-studied as exposures of the Lexington and Cincinnati Group rocks are easily accessible and numerous in the region between the Jessamine Dome and the Ohio River at Cincinnati. Nonetheless, this same region affords little access or exposure to the mid to lower Mohawkian strata, and no surficial exposure to the underlying Ashbyan rocks which are only accessible through cores and in subsurface mining operations. This study seeks to synthesize existing outcrop and subcrop investigations for the entire Chazy-Black River-Trenton group interval into a single reference document. These previous data are, at the same time, integrated with numerous new and unique stratigraphic observations afforded by the analysis of numerous rock cores through the Cincinnati, Lexington, and High Bridge Group rocks. Thus the important contributions of this study include detailing the approximate positions of key stage and sub-stage boundaries as applied elsewhere, as well as recognition of important distinct lithologic units and key contact intervals. These include the Ashbyan stage, and the Turinian, Rocklandian, Kirkfieldian, and Shermanian sub-stages of the Mohawkian Stage. The approximate equivalents of these intervals in the Nashville Dome region are also established based on biostratigraphic and lithostratigraphic correlation. Important biostratigraphic intervals are also detailed especially for the upper High Bridge to lower Lexington Group interval. Combined with lithologic indicators, these provide important insights into the nature and timing of extinction events, as well as climatic and oceanographic changes that likely resulted from a combination of sea-level oscillations and tectonic influences. These data are subsequently used to construct a detailed

stratigraphic framework of 13 time slice increments for comparison to other regions across the GACB.

INTRODUCTION:

Upper Ordovician rocks (460.5 to 449 mya) are well-exposed along the Cincinnati Arch and provide opportunities to investigate the stratigraphic history of the Taconic Orogeny. The Cincinnati Arch is a relatively broad (>160 km), but elongate (~800 km-long) structural province with a north-south orientation extending from Ohio, through central Kentucky into west-central Tennessee and northernmost Alabama. The arch is a broad, antiformal feature that extends to basement and has a series of intervening uplifts and sags or saddles along its axis. The central part of the Cincinnati Arch is commonly divided into the Nashville Dome and the Lexington or Jessamine Dome with a distinct region located in between referred to as the Cumberland Saddle which lies near the Tennessee-Kentucky border (**Figure 1**). The arch is particularly important in that that it separates the Appalachian and Illinois Basins to the east and west respectively, and with its lateral northward extensions (the Findlay and Kankakee Arches) the Cincinnati Arch also helps form the southern terminus of the Michigan Basin. The arch has been considered to have been a relatively stable portion of the North American continent although it appears to have been intermittently active during the Paleozoic and was an important feature in the Taconic Orogeny.

The Jessamine Dome, like other areas of eastern North America, has a relatively continuous succession of Upper Ordovician strata that have long been equated to the Chazy, Black River and Trenton Groups of New York State. As with exposures in the Appalachian Basin in central Pennsylvania, and central and eastern New York, strata of the Jessamine Dome also record the biotic and lithologic transitions associated with the tectonic activation of

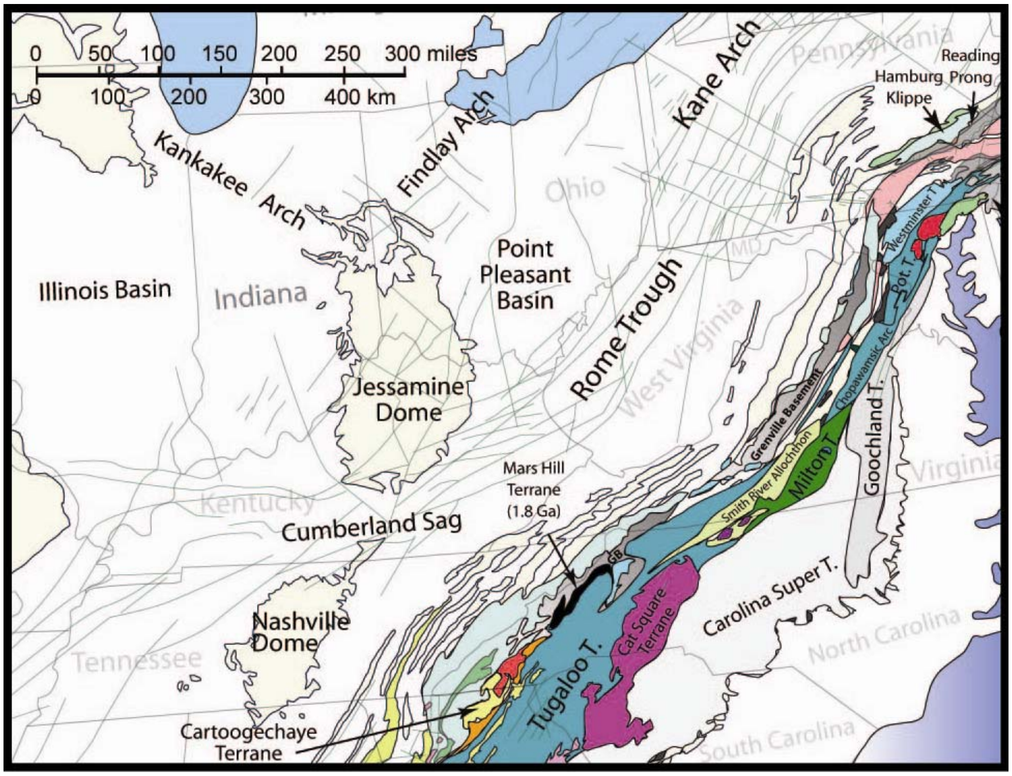


Figure 1: Ordovician outcrop areas (light yellow) of the Cincinnati Arch to the central Appalachian region. Also shown are fault and lineament trends as well as major tectonic elements of the Appalachian region. These include a number of exotic terranes, allochthons, basement uplifts, and potential Ordovician volcanic arcs. Most of which appear to have been emplaced during the early to middle Paleozoic. (Abbreviations: T = Terrane, Pot. T = Poteet Terrane, GB = Grenville Basement).

the eastern margin of Laurentia during the Taconic Orogeny. Given the position of the Cincinnati Arch in the Ordovician (roughly 300 km inboard of the cratonic margin), the region was far removed from the collision zone and was relatively stable and did not itself undergo major collision-induced subsidence. Nonetheless, the region did experience some far-field tectonic events as discussed elsewhere herein and by other authors (Holland & Patzkowsky, 1998; Ettensohn et al., 2002, Ettensohn et al., 2004).

In order to evaluate the timing and spatial development of the Cincinnati Arch in the context of other GACB regions during the onset of the Taconic Orogeny, the discussion here focuses on a stratigraphic review of the Upper Ordovician in the Jessamine Dome region. Emphasis here is on establishing the lithostratigraphic context of correlative biostratigraphic zones, various event horizons, and, as possible, the relative positions of important contacts that correlate with the CBRT interval of the type region in New York and Ontario. Establishing these contacts is especially important as the Cincinnatian Series of the arch region appears in stratigraphic succession above the Mohawkian Series of the type region and it is critical that the boundaries and internal stages are defined more accurately for biostratigraphic and evolutionary studies, as well as understanding their implications of the Taconic Orogeny on the GACB.

Stratigraphic Framework& Review of Previous Work

Near the close of the 19th century, rocks exposed along the axis of the Cincinnati Arch in the central to southern Bluegrass Region were recognized as the oldest strata exposed in Kentucky. Due to faulting and erosion of the central portion of the Jessamine Dome, rock strata equivalent to the Black River Limestones of New York State, have been brought to the surface along the Kentucky River in the vicinity of the Kentucky River Fault System. They are well-exposed from the position of Boonesborough, KY downstream to at least the position of the capitol city of Frankfort where the gentle westward dip takes these rocks below the ground surface (**figure 2**). Since the work of Campbell (1898), who first employed the terms Lexington and High Bridge Limestones, a number of workers have investigated the succession of rocks lying below those of the well-studied Cincinnati region downward to the lowest levels exposed in the Kentucky River Region. In addition a number of workers have also published some details of subsurface units below the level of the High Bridge Limestones. Important workers have included A.M. Miller (1905, 1913, 1915, 1919) who first sub-divided the Lexington & High Bridge Limestones on lithologic grounds, as well as Foerste (1906, 1909, 1913a, b, c) who

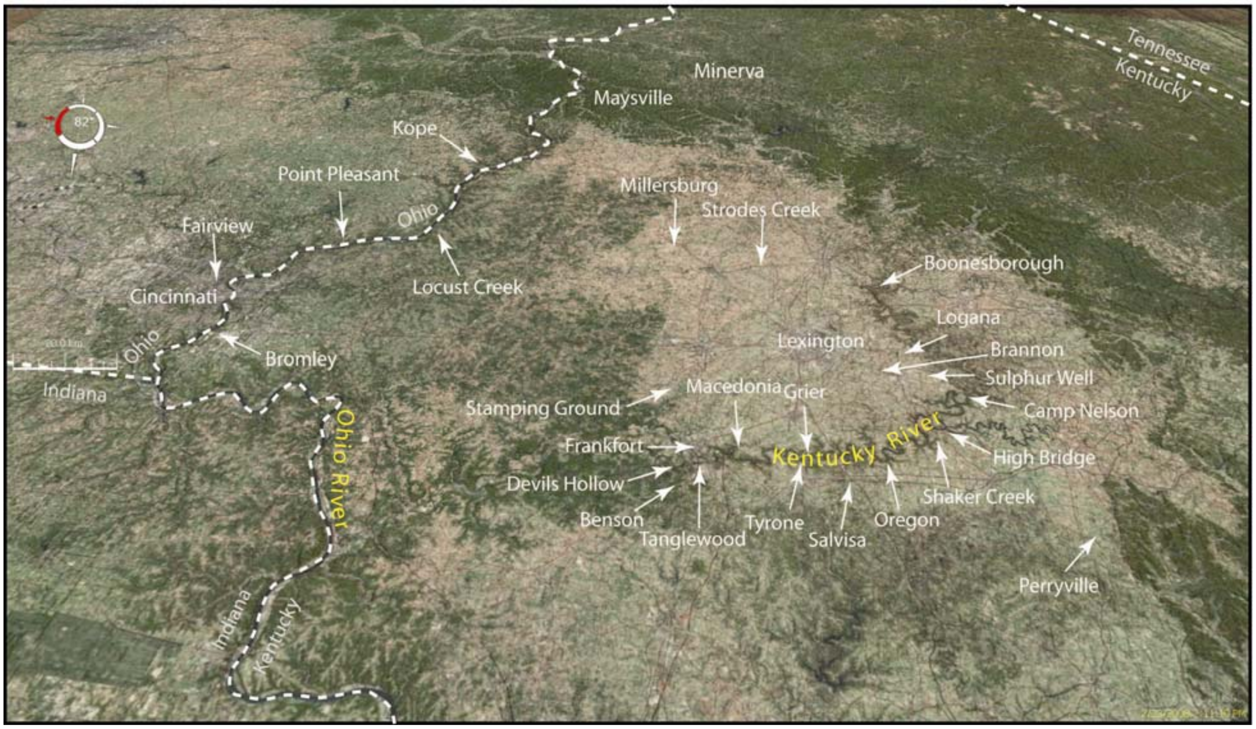


Figure 2: Type outcrop localities for stratigraphic intervals of the Upper Ordovician of the Cincinnati Arch – Jessamine Dome Region. Perspective view map produced using images (topographic & satellite) obtained from NASA World Wind digital globe software.

helped establish biostratigraphic frameworks, and McFarlan and colleagues (1931, 1938, 1943, 1948) who further worked to refine the lithostratigraphic and biostratigraphic succession of the Jessamine Dome region (**Figure 3**). USGS studies in the 1960s and 1970s (Black et al., 1965; and Pojeta et al., 1979) further investigated stratigraphic units in the region and replaced earlier syntheses through the abandonment and amalgamation of numerous stratigraphic units with re-definition of earlier terms. This work has been favored and used in recent analyses including those of Ettensohn (2002) and Ettensohn and colleagues (2002, 2004) who have recognized the role of distant tectonic activity in producing lithologic heterogeneities and unconformities along the axis of the Cincinnati Arch. One consequence of this early work has been establishment of the approximate stratigraphic age of these rocks in comparison to the type region (**Figure 4**). Disagreement concerning specific internal stage-level boundaries has been prevalent and has resulted in numerous chronostratigraphic assessments – most of which are in slight disagreement

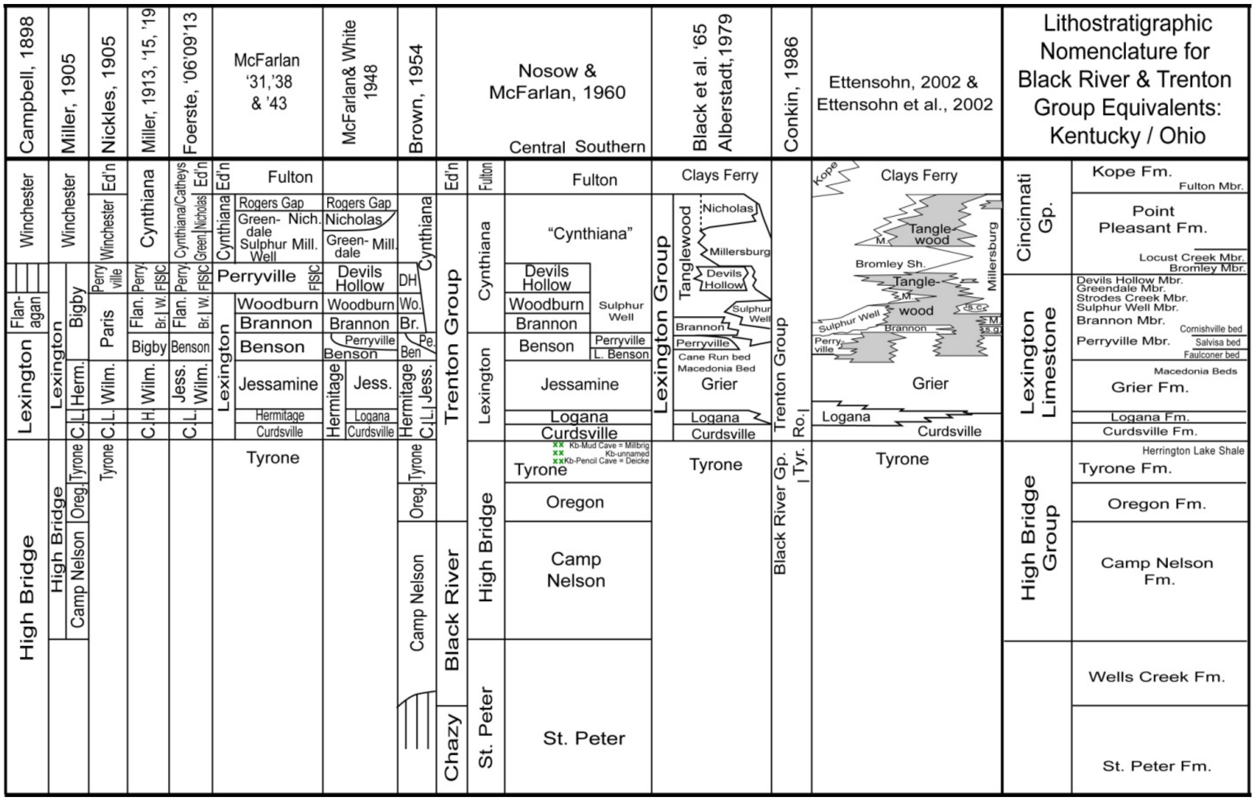


Figure 3: Temporal development of stratigraphic nomenclature for the lowest succession of rocks exposed in the Cincinnati Arch region versus stratigraphic nomenclature used by Brett et al., 2004, and herein. The lowest strata actually exposed in the Kentucky River region are limited to the interval above the lower Camp Nelson Formation.

with each other and in need of updating.

In the Jessamine Dome outcrop region, the CBRT equivalents are not fully exposed so previous estimates of thickness are based on combined outcrop and core data for this interval. As defined by Miller (1905) and later supported by Cressman (1964) and Cressman & Noger (1976), the High Bridge Group is not completely exposed but is considered to be composed of three formations. Nonetheless, in outcrop exposures the interval is about 134 meters thick and in the subsurface, the High Bridge is extended to as much as 198 meters. In these studies the High Bridge Group is defined to include the interval from the top of the Tyrone Limestone down to the base of the Wells Creek Dolomite above the Knox Group (Wolcott et al., 1972). As there is very little evidence for the St. Peter Sandstone equivalent in this region, the argillaceous base of the Wells Creek is considered the base of the High Bridge. Work in southern Indiana on the

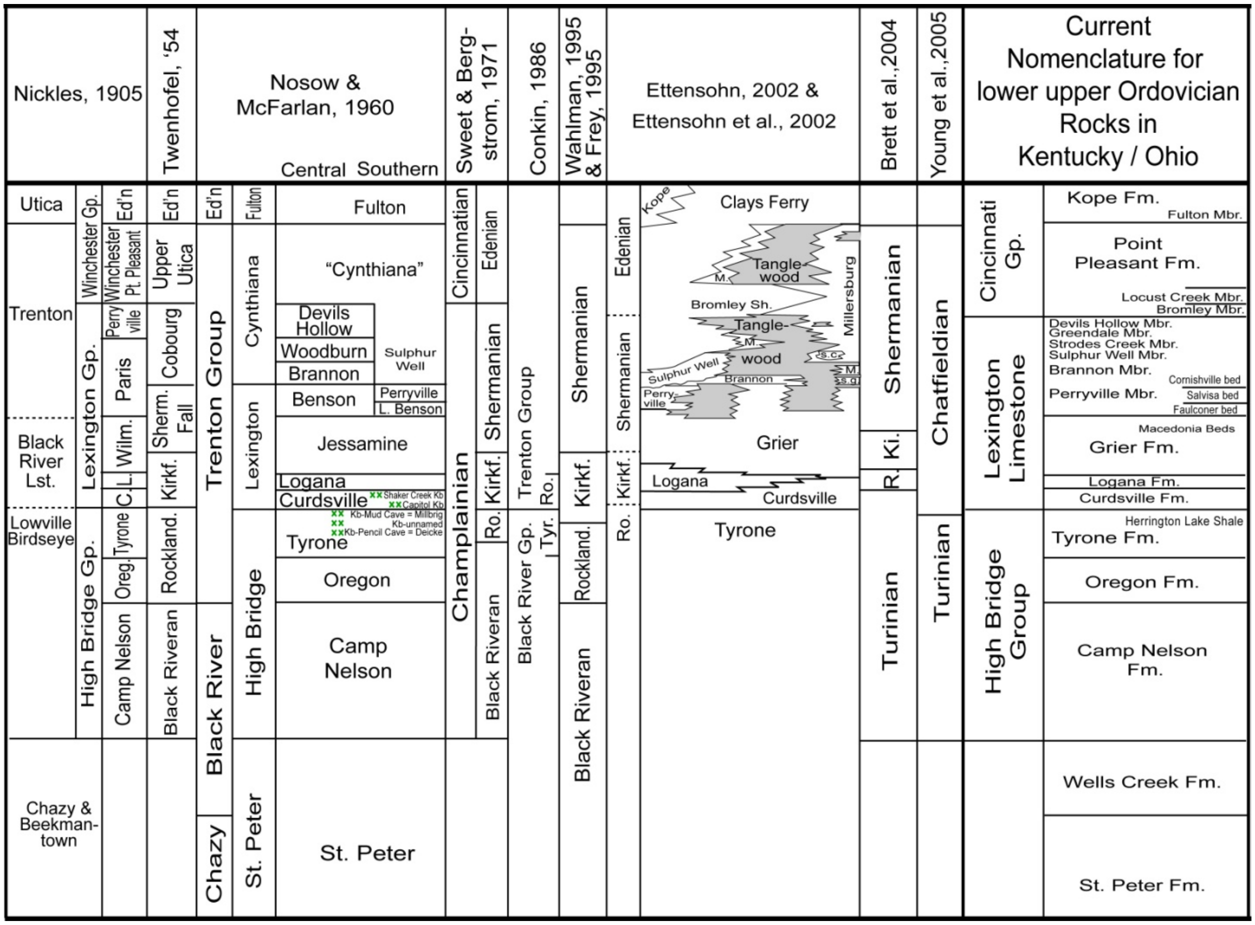


Figure 4: Review of previous chronostratigraphic assessments for the High Bridge and Lexington Groups of the Cincinnati Arch region.

west side of the Cincinnati Arch, shows a thickness of 152 meters for the Black River equivalent interval, with additional thickness of Chazy equivalents (Ancell Group) ranging up to 137 meters to the base of the St. Peter Sandstone (Droste et al, 1982). Thus in southern Indiana in the Illinois Basin the CBR interval ranges up to a maximum of about 289 meters. To the east of the Cincinnati Arch, in central Pennsylvania the CBR equivalent interval is estimated to be about 166 meters thick.

In the Jessamine Dome region, the overlying Lexington Limestone, up to the base of the Clays Ferry Formation, ranges in thickness upward to a maximum of about 98 meters (Wahlman, 1992). In Indiana the Trenton/Lexington equivalent has a maximum thickness of about 81 meters in northeastern Indiana and is considerably thicker in southeastern Indiana where it transitions into the Sebree Trough area before thickening again onto the Cincinnati Arch. To the east of the Cincinnati Arch in central Pennsylvania the Trenton equivalent is nearly double the thickness of the Cincinnati Arch – Illinois Basin region and approaches 183 meters thick.

Thus in total the High Bridge to Lexington interval on the Cincinnati Arch is recorded by upwards of 300 meters of strata deposited during the early phases of the Tippecanoe transgression. Thicknesses for this entire interval increases both to the west and east. On the western flanks of the arch in southern Indiana the maximum thickness approaches 370 meters inclusive of about 43 meters of St. Peter sandstone, whereas in central Pennsylvania to the east of the arch, the CBRT equivalent strata are somewhat less at approximately 349 meters thick. Thus, although the thickest interval of the entire CBRT occurs to the west of the Cincinnati Arch (370 meters) and the thinnest interval occurs on the arch, this pattern is attributable to the greater thickness of Chazy-Black River (CBR) equivalent strata only. Moreover, if the thickness of the St. Peter sandstone equivalents is excluded, this region has a carbonate-dominated thickness of about 327 meters which is more comparable to values in Pennsylvania. In contrast, the thicknesses of Trenton equivalents show a reversed trend from west to east. In Indiana and the Cincinnati Arch there is significantly less Trenton (81 and 98 meters respectively) than in Pennsylvania (183 meters).

Overall the changes in thicknesses and therefore accumulation rates suggest that deposition both on the Cincinnati Arch and to the east in central Pennsylvania was somewhat less than depositional rates in the Illinois Basin. However, it is clear that the Cincinnati Arch and Illinois Basin witnessed greater accumulation rates during the CBR interval than did the central Pennsylvania region, whereas during deposition of the Trenton, this pattern reversed. In the Trenton, significant carbonate accumulation (with admixed siliciclastic contributions from the Taconic Highlands) occurred in Pennsylvania owing to the impact of the Taconic Orogeny.

Stratigraphic boundaries of the study interval: lower contact – basal Ashbyan

As the basal High Bridge Group is not exposed and is characteristically sparsely fossiliferous it has received significantly less study than the overlying Lexington Limestone. Therefore, the lowermost lithostratigraphic and chronostratigraphic boundary of the CBRT interval has not been placed in the Cincinnati Arch region in outcrop. In the subsurface, Wolcott and colleagues (1972) favored the base of the High Bridge to be placed at the base of the Wells Creek Formation; however, most workers employ the original definition for the High Bridge Group (after Miller, 1905). In the subsurface of southernmost Ohio and northern Kentucky, Stith (1979) placed the base of the Black River Group (and the High Bridge Group) at the base of his "lower argillaceous member" which he correlated to be the base of the Camp Nelson (**figure 5**). As few studies have established age equivalency for underlying units, Stith (1979) reported age estimates for the strata underlying the High Bridge Group based on correlations in Indiana. In southern Indiana the Wells Creek and underlying St. Peter (Ancell Group) have been considered to be equivalents of the Chazy based on conodont analysis (Sweet and Bergström, 1976; Droste et al., 1982). The base of these units has long been equated with the position of the Knox Unconformity.

In the Cincinnati Arch region, the contact is less distinct without the recognition of the St. Peter Sandstone which can be absent in some areas. However Stith (1979) recognized the base of the Chazy-Wells Creek using well logs (**see figure 5**). In northern Kentucky, gamma-ray

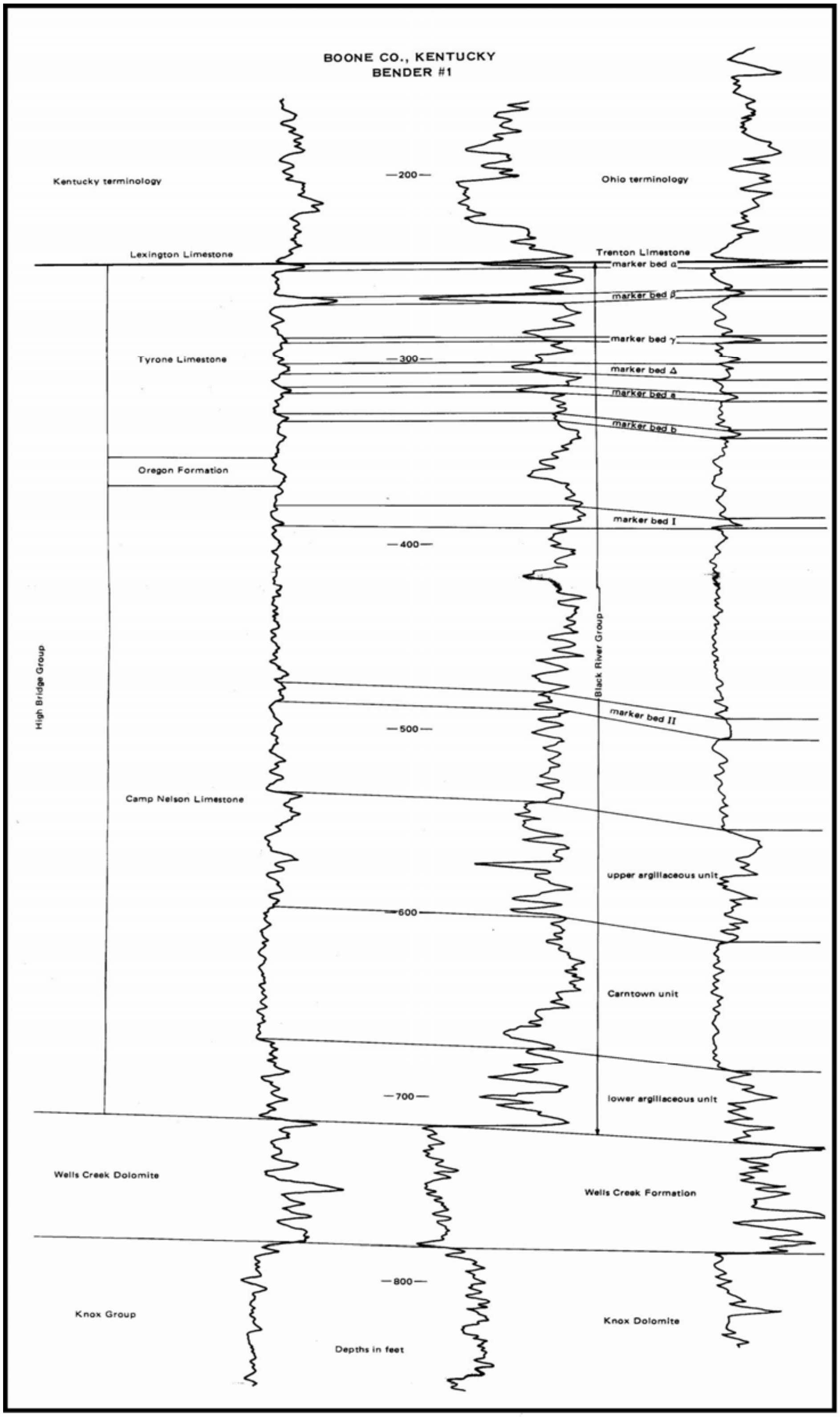


Figure 5: Gamma Ray and Neutron Logs representative of the CBR interval of the northern Kentucky – southern Ohio region. Image after Stith (1979).

signatures show a pronounced increase in API values at the top of the Knox Dolomite (base of the Wells Creek) which is attributed to the influx of siliciclastic clays above the Knox. Neutron logs show a minor drop in values. Gamma-ray signatures remain somewhat high throughout the Wells Creek and into the base of the overlying “lower argillaceous unit” which Stith considers to be Black River – but may in fact still be Chazy equivalent. Neutron log values, however, show a sharp kick to higher values at the top of the Wells Creek Formation and at the base of the “lower argillaceous unit.” In this study, the assessment of Stith is followed with the lowest contact of the CBR interval established at the base of the St. Peter where present, or at the base of the overlying Wells Creek Formation. Based on the work of Droste and colleagues (1982) conodonts from the basal St. Peter are Chazy or “Ashbyan” in age.

<u>Stratigraphic boundaries of the study interval: upper contact – upper Shermanian</u>

The upper contact of the CBRT interval, although the subject of numerous previous investigations, is still poorly defined and is not consistently applied along the axis of the Cincinnati Arch from southwestern Ohio into central Kentucky. This is in part due to complicated stratigraphic nomenclature and somewhat parochial differences in the application of stratigraphic names. In northern Kentucky to southern Ohio, the pronounced contact at the base of the “Eden Shales” (now base of the Fulton beds of the Kope Formation) and the top of the underlying Point Pleasant was established by Nickles (1905) as synonymous with the Trenton-Utica contact of the type region (**see Figure 4**). Thus most subsequent workers (i.e. Lattman, 1954; Nosow and McFarlan, 1960) have considered the Point Pleasant and underlying units to be contained within the Trenton Group and thus Mohawkian in age. Moreover, beds overlying the

Point Pleasant and beginning at the base of the Fulton were by definition Edenian in age and representative of the Cincinnatian Series.

Farther south, in the Jessamine Dome region, McFarlan and colleagues (1931, 1938, 1943, 1948, 1960) recognized equivalent strata including the Fulton Shale and the underlying Rogers Gap and Nicholas members of the Cynthiana Formation (**see Figure 2**). These latter units were considered equivalents of the Point Pleasant and also included in the Trenton Group (Cressman, 1973). Subsequent work by Weir and Greene (1965) in Fayette County, Kentucky failed to recognize the same pronounced contact separating the Cynthiana from the overlying beds. Thus these authors proposed the term Clays Ferry to include the Cynthiana and overlying strata through to the base of the Garrard Siltstone. Along with Black and colleagues (1965), these authors implemented a number of changes in the lithostratigraphic nomenclature. These included abandonment of Foerste's (1906, 1909, 1924) Cynthiana Formation in favor of a new series of terms based entirely on lithofacies. Black and colleagues introduced the term Tanglewood Limestone for the coarser-grained limestone facies through much of the interval previously considered to be Cynthiana and established lateral equivalents including a number of tongues and lentils. The Tanglewood was interpreted to transition laterally (to the east and west) into somewhat more siliciclastic-rich intervals of the Millersburg. Thus Black and colleagues (1965) expanded the Lexington Limestone upward to the top of the Nicholas Limestone (the coarse-grained facies equivalent to the upper Pt. Pleasant Foramtion at the top of the Mohawkian) and included all of the underlying units in their modified Lexington Group. The upper contact of the Lexington was placed at the base of the Clays Ferry Limestone of Weir and colleagues (1965). However, given the significantly lowered siliciclastic (shale) component of overlying strata in this region, and assumed diachroneity of lithologic units, the contact between the Clays Ferry and the underlying Tanglewood has been interpreted as relatively gradational and diachronous over much of the Jessamine Dome (Ettensohn, 2002). Under the Weir and colleagues model (1965), the Clays Ferry was expanded northward to include the Point Pleasant as a tongue of the Clays Ferry. This assessment thus places the Tanglewood-Clays Ferry contact below the level of the Point Pleasant-Kope contact which is considered to be the equivalent of the top Trenton contact. As a result the top contact of the Trenton has been obscured within the complex stratigraphic nomenclature. Moreover, macrofaunal biostratigraphic analyses have been unable to resolve these difficulties, but simply confirm that the Clays Ferry is likely Shermanian to Edenian in age (Wahlman, 1992).

Nonetheless, work by Mitchell and Bergström (1991) using graptolite and conodont biostratigraphy has helped to recognize key biostratigraphic zones between the Cincinnati Arch and the type Mohawkian region. In New York, the top of the Trenton at Trenton Falls is placed at the top of the Steuben Limestone where it is unconformably overlain by the Indian Castle Shale (Baird & Brett, 2002). As shown by Mitchell and Bergström (1991) the approximate position of the Steuben-Indian Castle Shale contact falls above the base of the *C. spiniferus* graptolite zone and below the base of the *G. pygmaeus* zone. In Ohio, these authors recognized the Point Pleasant – Kope lithostratigraphic contact to also lie above the base of the *C. spiniferus* zone and below the *G. pygmaeus* zone boundary. Based on these assessments, event stratigraphy, and sequence stratigraphic correlations from central Kentucky northward to the Cincinnati region, Brett and colleagues (2004) established the top contact of the Trenton Group at Trenton Falls to be synonymous with the Point Pleasant-Kope contact. Nonetheless, it is recognized that the top of the Trenton Group in northwestern New York and Ontario is somewhat younger with the top of the Trenton assigned at the top of the Hillier Limestone

(Lehman et al., 1994). The top of the Hillier is approximately coincident with the *C. spiniferus* – *G. pygmaeus* graptolite zone boundary which in Ohio and northern Kentucky occurs in the lower to middle Kope Formation. Given the predominance and definitive lithostratigraphic base of the Edenian at the base of the Kope – it is suggested that the Pt. Pleasant-Kope contact is not only the top of the Lexington Group (as defined by Black et al., 1965), but is also the top of the Trenton Group equivalent as suggested by Nosow and McFarlan (1960).

<u>Internal Stratigraphic Boundaries: Stage – and Substage boundaries</u>

Early attempts at establishing specific age equivalents for stages between the type region and the Cincinnati Arch region within the CBRT interval were limited to investigations of the early 1900's when New York was the standard for comparison and establishing correlations. Subsequent work focused on more local correlations with strata exposed primarily in the Tennessee region. The first significant attempt to establish stage-level classifications was through the large compendium of correlations for the Ordovician of North America (Twenhofel et al., 1954). This large compendium produced the first North American-wide correlation chart with over seventy localities across the continent for the Ordovician. In this analysis, Twenhofel and colleagues established a framework of correlations for the specific purpose of summarizing existing stratigraphic data and to provide for a framework within which to continue to refine and develop correlations across the region.

In the Cincinnati Arch region, the High Bridge, Lexington, and Cynthiana Limestones (up to the level of the Fulton) were considered equivalents of the BRT of New York. As the lower High Bridge was unexposed in Kentucky – no Chazy strata were recognized in their study. Nonetheless, stage boundaries from the New York Mohawkian Standard were placed relative to strata of the Jessamine Dome region. Twenhofel and colleagues (1954) recognized the Black River and Trentonian stages (the latter of which was subdivided into the Rockland, Kirkfield, Sherman Falls, Cobourg, and tentative “Upper Utica” sub-stages) below the level of the Edenian (**see Figure 4**). Subsequent biostratigraphic work on conodonts by Sweet & Bergstrom (1971) and Sweet (1984) resulted in minor modifications of these age assessments – the details of which are discussed below.

Ashbyan Stage

As Chazy strata are unexposed in the Jessamine Dome region, details of this stratigraphic interval have been limited except for in the subsurface. As the base of the Upper Ordovician has been placed at the base of the *N. gracilis* graptolite zone, recognition of this specific interval has not been established in the Cincinnati Arch region. Nonetheless, this graptolite zone roughly corresponds to a position in the middle of both the *Pygodus anserinus* and *Cahabagnathus sweeti* conodont zones (Webby et al., 2004; **Figure 6**). Data presented by Votaw (1972) and Sweet (1984) shows much of the exposed sections of the High Bridge Group to overly the *C. sweeti* zone. However the basal Camp Nelson and the immediately underlying upper “Knox” interval contain conodonts of the *C. sweeti* zone. These workers did not formerly recognize the Wells Creek interval and likely included it with the underlying Knox. However as suggested by Stith (1979) in support of work by Calvert (1962) and Carpenter (1965) and based on the work of Droste and colleagues (1982) in southern Indiana, the Wells Creek and underlying St. Peter contain conodonts indicating a Chazy or Ashbyan age. Thus herein the base of the Ashbyan is placed at the unconformable base of the St. Peter or where that unit is not present owing to non-deposition and/or erosion along the Cincinnati Arch, the base is established at the base of the slightly younger Wells Creek.

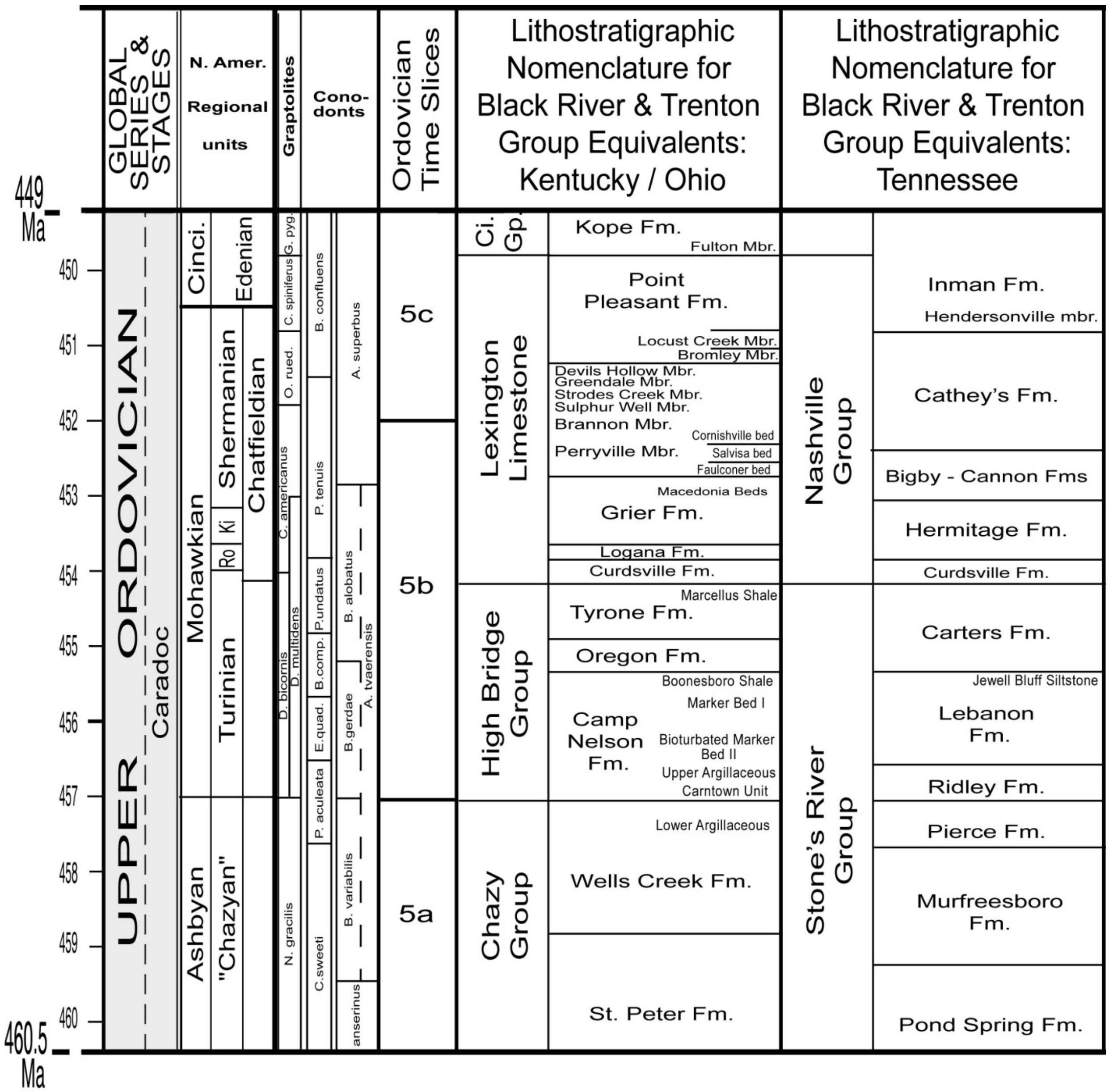


Figure 6: Biostratigraphy and lithostratigraphic nomenclature, as applied herein for the Cincinnati Arch region including the Jessamine and Nashville Domes. Where present, the base of the St. Peter represents the base of the Upper Ordovician strata in the Cincinnati Arch region.

The upper boundary of the Ashbyan is established at the base of the *Diplograptus bicornis* graptolite zone and at the base of the *Baltoniodus gerdae* conodont subzone of the *Amorphognathus tvaerensis* zone (Webby et al., 2004). The top of the Ashbyan is also now defined just below the top of the *Plectodina aculeata* conodont zone which Sweet (1984) used to bracket the Ashbyan. In the Cincinnati Arch region, Votaw (1972) and Sweet (1984) indicated that at least the lower Camp Nelson interval (Carntown Unit of Stith, 1979) was representative of the *P. aculeata* Zone. As the *P. aculeata* zone brackets the Ashbyan-Turinian boundary, it is surmised that the upper contact of the Ashbyan occurs within the base of the Camp Nelson. Based on other correlations established herein it is likely that the Ashbyan top contact occurs within the interval above the “lower argillaceous interval” of Stith and below his “upper argillaceous” zone. In the vicinity of the Kentucky River near the type section at Camp Nelson, both the uppermost Carntown through the “upper argillaceous” interval appear to be exposed, but a specific contact is not yet recognized. If it is present it is likely at the very base of the exposed sections of the Camp Nelson or just below the surface exposures.

Turinian Stage

The Turinian Stage of the New York standard section was established to include the interval originally included within what had been classified as the Black Riveran. As the Black River Group and therefore the Turinian Stage has its base drawn at the level of the lowermost Pamelia Formation where in the Champlain Valley its sits immediately above the Chazy Limestones with only minor evidence for unconformity. In New York, the Turinian strata of the Black River Group are documented to contain conodonts from the uppermost *P. aculeata* zone the *Erismodus quadridactylus,* the *Belodina compressa,* and *P. undatus* zones (Sweet, 1984). In New York, the conodont assemblage also contains forms that define the *Baltoniodus gerdae* and *Baltoniodus alobatus* North Atlantic assemblages, with the base of the *B. gerdae* zone located at the approximate level of the Chazy-Black River contact (Sweet, 1988).

In the Cincinnati Arch region, the Camp Nelson, Oregon, and Tyrone intervals of the High Bridge group contain conodonts characteristic of the same mid-continent zones recognized in the northeast. That is the High Bridge Group in its exposure region contains *P. aculeata, E. quadridactylus, B. compressa*, and *P. undatus* (Votaw, 1972; Sweet, 1984; Leslie & Bergström, 1995). Given the correspondence of biofacies as well as lithologic similarities as outlined by previous authors, the High Bridge Group of central Kentucky is approximately equivalent to much of the Black River Group and can be largely ascribed to the Turinian. As mentioned the lower contact with the Ashbyan likely occurs within the Carntown unit of Stith (1979).

The upper contact of the Black Riveran (or Turinian) in Kentucky has been the subject of much debate as shown in **figure 4**. Beginning with Twenhofel (1954) the uppermost "Black Riveran" was capped at the upper contact of the Camp Nelson, with the overlying Oregon and Tyrone considered Rocklandian in age. This was in contrast to original suggestions that the Tyrone of the Cincinnati Arch was actually the Lowville equivalent (Nickles, 1905). Subsequent work including the work of Sweet & Bergstrom (1971) based on conodonts and work by Conkin (1986) based on unconformities and K-bentonite correlations pushed the upper contact of the "Black Riveran" or Turinian upward to just below the top of the Tyrone and to the top of the Tyrone respectively. More recently Leslie and Bergstrom (1995) proposed that the top of the Turinian be established at the level of the Millbrig K-bentonite which when present in the Jessamine Dome occurs just below the top of the Tyrone. Alternatively, when the K-bentonite is removed by unconformity the Turinian-Chatfieldian boundary is established at the position of the distinct Tyrone-Curdsville (Lexington Limestone) contact. Nonetheless, with recognition of the Millbrig K-bentonite in New York just below the top of the Black River Group (at the base of the Watertown Limestone) it is suggested that the type Turinian projects upward some distance above the Millbrig and into the overlying Curdsville Limestone. Thus herein, the top of the Turinian is established to a level within the middle of the Curdsville Limestone where depositional facies begin to show the rapid deepening-upward patterns typical of the Turinian-Rocklandian transition of New York. Based on correlations of the GICE, K-bentonites, and sequence stratigraphy, the top of the Turinian should be established at the approximate level of the *Phragmodus undatus – Plectodina tenuis* conodont zone boundary as discussed previously.

Chatfieldian Stage

The Chatfieldian Stage was proposed by Leslie & Bergström (1995) as the upper stage of the Mohawkian Series. Based on the historic difficulties and conflicts associated with recognition of the type "Trentonian" stages in the Jessamine Dome region, it was proposed in order to provide a chronostratigraphic reference that could be correlated and recognized across the GACB. Although it has been used by some recent workers, this term has not generally been applied by most workers in the Jessamine Dome region and historic chronostratigraphic assessments continue to be used in the literature (i.e. see Wahlman, 1992; Ettensohn, 2002). As the Rocklandian, Kirkfieldian, and Shermanian stages are entrenched in the literature there are a number of reasons to continue to use these terms, although their application in Kentucky and Ohio has to be reconsidered and updated. Thus here the original stages for the "Trentonian" are reevaluated although it is recognized that another appropriate stage name is needed to replace the term "Trentonian."

Rocklandian, Kirkfieldian, and Shermanian Stages

The stages proposed by Kay (1943, 1944), as discussed previously for the New York-Ontario type region, were defined on both lithologic and biostratigraphic grounds. A number of important brachiopod, mollusk, echinoderm, and trilobite taxa have been employed in establishing the relative boundaries of these units. In addition, graptolites and conodonts have been used in the type region to help refine the boundaries of these stratigraphic stages (Schopf, 1966; Barnes, 1969; Riva, 1974; Sweet, 1984; Goldman et al., 1994) in an effort to enable regional correlation. In New York the base of the Rocklandian is relatively synchronous with the *P. undatus – P. tenuis* conodont zone boundary and lies within the middle of the upper *A. Tvaerensis* subzone (*B. alobatus* Zone). In addition, as described by Webby and colleagues (2004), the *Diplograptus bicornis- Corynoides americanus* graptolite zone boundary is also approximately coincident with the base of the Rocklandian while the top of the *D. multidens* Zone is thought to be approximately equivalent of the top of the Kirkfieldian. In New York and Ontario, Sweet (1984) suggested that the boundary between the Turinian and the Rocklandian could be recognized by the relatively dramatic decline to near extirpation of fibrous conodont forms in the region. Although likely tied to water depth changes, subsequent Rocklandian conodont faunas became dominated by the long ranging *P. undatus* (~60-80% of the forms) and a few forms included in the genera *Belodina*, *Panderodus*, and *Plectodina*. The Rocklandian-Kirkfieldian boundary was indistinguishable; however the top of the Kirkfieldian and base of the Shermanian was noticed by the shift to nearly 95-100% dominance of the fauna by *P. undatus*. *Belodina*, *Panderodus*, and *Plectodina* did not return to significant abundance until the upper Trenton during what was referred to as the CoboURGian (now upper Shermanian) when these forms replaced *P. undatus* almost entirely.

In central Kentucky within the Lexington Limestone, very few graptolites have been identified and thus they are not useful for biostratigraphic and chronostratigraphic analysis in this region. Therefore age-based correlation with the type region has relied upon macrofaunal taxa as well as available conodont biostratigraphic data, and most recently the Millbrig K-bentonite. Historically, Bassler (1932) and other workers considered the Tyrone as Lowville, as did previous workers including Nickles (1905). However, work by Twenhofel and colleagues (1954) indicated that the Tyrone was younger than the upper Black River as the unit was a correlative of the Moccasin of Virginia based on K-bentonite correlations. Due to its siliciclastic content, the Moccasin was always considered to be Trenton. However, it was recognized that the faunas of the Tyrone were most like those of the Lowville. Age estimates were also supported by the recognition of a diverse echinoderm fauna in the Curdsville Limestone through the base of the Jessamine (Grier Limestone). The presence of these echinoderms has been reported by numerous workers including Nosow and McFarlan (1960), Bell (1979), Branstrator (1979), etc.. Similar echinoderm faunas are noted from the Lebanon Limestone of Tennessee as well as the Kirkfieldian of Ontario. As the Lebanon was correlated with a portion of the underlying High Bridge Group, the echinoderms have been used to support a Kirkfield equivalency for the Curdsville through lower Grier interval.

In terms of conodonts, numerous fibrous conodont forms have been recognized in the High Bridge Group (Votaw, 1972, Sweet 1984). These give way upward to a thin interval at the base of the Curdsville dominated by *Plectodina*. Immediately above, the rest of the Curdsville shows that *P. undatus* becomes the dominant form and maintains its abundance upward into the Lexington Limestone (Sweet, 1979). Near the middle of the Lexington, conodont associations again become dominated by *Plectodina* related forms. As in New York the dominance of *P. undatus* suggested that the Curdsville and the Grier were equivalent to the Rocklandian, Kirkfieldian to lowermost Shermanian. The increased abundance of *Plectodina* in the middle to upper Lexington supported previous estimates of a Shermanian age as is suggested here. Sweet thus (1979) favored a position for the Kirkfieldian-Shermanian boundary to lie within the Grier a position that has been maintained by most workers including Wahlman (1992), Frey (1995), and

Ettensohn (2002). In terms of the Rocklandian and Kirkfieldian boundaries, the uppermost Tyrone to lowermost Curdsville has been considered as Rocklandian including the interval within which the Deicke and Millbrig K-bentonites occur (Sweet, 1984; Kolata et al., 1996). The majority of the Rocklandian has thus been considered to occur within the stratigraphic gap between the High Bridge and Lexington Limestones where up to 4 meters of strata have been suggested as missing (Cressman, 1973). Recent work suggests that the correlation of the Tyrone to lower Curdsville interval as Rocklandian, and the upper Curdsville to Grier as Kirkfieldian is problematic and is discussed below.

The uppermost contact of the Shermanian, although considered originally to correlate with the top of the Benson – Perryville interval (Twenhofel, 1954, Nosow & McFarlan, 1960), has been extended upward on the suggestion of Ross (1982). Most workers today consider the top of the Shermanian to coincide with the top of the “Cynthiana” of earlier workers, and the top of the Tanglewood Limestone (Ettensohn, 2002). Thus, the top of the Shermanian is located at the contact between the top of the Point Pleasant Limestone and the overlying Kope Formation (Bergström & Mitchell, 1990; Mitchell & Bergström, 1991). This evaluation is established based on the projected last occurrence of *Orthograptus ruedemanni* from the Sebree Trough equivalents of the Point Pleasant and the occurrence of *C. spiniferus* as well as the conodont *Belodina confluences* within the Point Pleasant. The overlying Kope Formation witnesses the end of the *C. spiniferus* and *B. confluens* zones well above the contact and within the Edenian Stage of the Cincinnatian Series.

Re-investigation of chronostratigraphic zones in the Cincinnati Arch

Recent recognition of the Guttenberg Isotopic Carbon Excursion (GICE) in the Logana to lower Grier and in the Napanee to lower Kings Falls of New York (as reported by Young et al., 2005 & Barta, 2004) along with identification of the Millbrig K-bentonite (Mitchell et al., 2004) below the excursion in both regions helps establish the position of the Rocklandian and Kirkfieldian outside of New York State. In the Cincinnati Arch region, the position of the Millbrig has long been established at a level just below the Curdsville Limestone. With resolution of stratigraphic nomenclature as suggested herein (see chapter 2, &3), the Millbrig is equivalent to a K-bentonite located below the sequence boundary at the base of the Watertown Formation in New York. Moreover, in New York the GICE is recognized some distance above the Millbrig interval within the Napanee Limestone.

Thus if the Millbrig occurs below the uppermost Turinian and the GICE occurs within the Rocklandian of the type region, this clearly indicates that the High Bridge Group is entirely Turinian in age with the base of the overlying Curdsville also lying within the Turinian stage. Moreover, the occurrence of the GICE within both the Napanee to lower Kings Falls and in the Logana to lower Grier suggests that the Logana to lower Grier is indeed not Kirkfieldian, but is Rocklandian in age. As suggested from sequence stratigraphy (Brett et al., 2004), and a re-evaluation of available conodont biostratigraphic data (Young et al., 2005) and macrofaunal biostratigraphic data as evaluated herein, it appears that former biostratigraphic assessments are permissive of a the relative age assessments from early workers (Nickles, 1905, Bassler, 1932). Thus with recognition of the Curdsville as mostly latest Turinian, and the Logana to lower Grier as Rocklandian, the position of the Kirkfieldian has to be considered to be equivalent to an interval within the overlying Grier member of the Lexington. This can be supported as the Grier is dominantly a bryozoan-echinoderm-dominated unit similar to the Kings Falls interval of New

York. Cressman (1973) reports a number of beds with large echinoderm stems preserved (likely a species of *Cleiocrinus*) up to five meters below the level of the Macedonia.

The distinct Macedonia interval near the top of the Grier, with its Logana-like facies some 20 meters above the base of the Grier and below the level of the Benson-Perryville interval, appears as an excellent candidate for the contact between the top of the Kirkfieldian and the base of the Shermanian. This stratigraphic position coincides approximately with the top of the Jessamine or Hermitage of early workers including Nosow and McFarlan (1960). This contact also occurs just above the approximate base of the Shermanian as established by Twenhofel and colleagues (1954) and supported by subsequent workers. Thus although the positions of the Rocklandian and Kirkfieldian have shifted upwards, the base of the Shermanian has only shifted a small distance upward to the level of a prominent and easily recognizable facies change to deeper water facies as occurs in the type region.

Chazy Group: General Description, Contacts & Distribution

In the Cincinnati Arch region, the lowest rocks to be documented included the Camp Nelson Formation of the High Bridge Group in the Jessamine Dome and the Murfreesboro Formation of the Stones River Group of the Nashville Dome (**see Figure 6**). Early workers recognized significant faunal and lithologic similarities between these two regions and many stratigraphic correlations have been established. As reported by Twenhofel and colleagues (1954), the High Bridge and Stones River Groups have been well-correlated and it is acknowledged that exposures in the Nashville Dome afford the opportunity to investigate strata below the level of the lower Camp Nelson. In the Nashville Dome, the lowest units exposed include the Murfreesboro and Pierce Limestones – both of which are correlated with the stratigraphic interval below that exposed in the Camp Nelson of the Kentucky River region. Thus, in the Jessamine Dome region, the lowest rocks exposed are above the level of the basal Chazy and in fact, it is likely that most of the exposed interval of the High Bridge Group is in fact Black River equivalent.

Thus, in the Jessamine Dome to southwestern Ohio region, the Chazy Group equivalents are generally limited to the St. Peter Formation, and the Wells Creek Formation – both of which are recognized in core and well-cuttings only. In the Jessamine Dome area, Matson (1909) recognized the St. Peter sandstone to lie approximately 30 meters or more below the base of the High Bridge Group and Fuller and Clapp (1912) identified the unit nearly 260 meters below surface rocks in the vicinity of Cincinnati. Calvert (1962 a, b) was the first to equate this interval with the Chazy of New York. Located above the St. Peter, the dolomitic limestones and dolomites of the Wells Creek Formation are also recognized. However, in many cases, where the St. Peter is absent below, the Wells Creek is often lumped with underlying "Knox" carbonates. This is similar to the scenario in central Pennsylvania where the Upper Bellefonte (Tea Creek member) is often included in the Beekmantown (Knox) although a thin interval referred to as the Dale Summit Sandstone intervenes. This thin, intermittently-recognized sandstone interval is likely the equivalent of the St. Peter making the overlying Tea Creek a Chazy equivalent unit.

Overall, few studies have investigated the nature of the Chazy equivalents in the Cincinnati Arch region and therefore details of these units are few. Nonetheless, as recognized herein following the work of Calvert (1962 a) and Carpenter (1965), some details are afforded which enable the base of the Chazy to be placed at the base of the St. Peter. The top of the

Chazy Group likely occurs within the argillaceous interval found in the interval immediately above the Wells Creek Formation. Thus overall the Chazy interval approaches a maximum of 38 meters and contains evidence for condensation and the influence of relict topography produced during development of the Knox Unconformity on the arch.

Chazy Group Lower Contact

The lower contact of the Chazy Group interval is considered to occur at the base of the St. Peter Sandstone on the flanks of the Cincinnati Arch. In southwestern Ohio and the northern Jessamine Dome region, Calvert (1962 a) and Carpenter (1965) have recognized significant and pronounced erosional topography on the Knox Dolomite Group. In this area, this topography results in the variable thickness of both St. Peter and the “Lower Dolomite Member of the Chazy Limestone” (aka Wells Creek / Murfreesboro). In this region, these units have been noted to range from absence to upwards of 40 meters thick where they are most easily recognized. Along the central axis of the arch in the central Jessamine Dome, where St. Peter has not been recognized as a pronounced sandstone unit, Calvert (1962 a) considered the contact between the Chazy and the underlying Knox at the position of a pronounced shaly interval in the base of what is now called the Wells Creek.

Recognized on well logs, owing to increased siliciclastic components, this interval shows a rapid increase in gamma-ray response and contains a basal conglomeratic zone (**see figure 5**). In cores this interval is recognized as an interval of fairly large clasts within an argillaceous dolomite. Nonetheless, well-rounded and frosted quartz grains are found in this interval and are often developed into lenses of sandstone usually less than 2 meters thick. Although not the dominant lithology – either these quartz-rich intervals are equivalents of the St. Peter or are slightly younger reworked equivalents. Nonetheless the prominent contact at the base of this unit is considered a sharp, erosional unconformity and is likely equivalent of the regionally well-exposed and well-developed Knox Unconformity (Carpenter, 1965).

Chazy Group Upper Contact

In the subsurface of southwestern Ohio, Calvert (1962 a) was the first to delineate Chazy equivalent strata. In his study, he diagnosed a tripartite Chazy Limestone with a "lower dolomite member," a "middle limestone member," and an "upper argillaceous (cherty) member" all below the level of his Black River Group. Carpenter (1965) and Stith (1979) recognized a similar succession and used gamma-ray and neutron logs for substantiating the contacts, but suggested that the upper members of Calvert's Chazy Limestone were in fact Black River. Thus, the lithologic change from argillaceous dolomitic carbonates to relatively pure fossiliferous limestones at the base of the "middle limestone" (of Calvert, 1962 a) was considered to be the upper contact of the Chazy by Carpenter (1965). Immediately below the relatively pure limestones of the middle limestone member, the interval shows a pronounced neutron log signature located between the Wells Creek Formation and the overlying "middle limestone." Termed the "lower argillaceous unit" by Stith (1979), this interval shows transitional facies between the Wells Creek below and the Black River above. Carpenter (1965) considered the top of this unit to be the top Chazy, acknowledging, however, the difficulty of placing a specific lithologic boundary in cored sections. Stith (1979) however, considered the base of this unit to be the top of Chazy rocks. Owing to significantly less dolostone and more typical micritic limestones in the lower argillaceous interval, Stith considered the lower argillaceous interval to be lithologically more similar to the overlying Black River/High Bridge Group interval and included it in the overlying unit. Thus there is still debate as to the position of the true CBR contact. Nonetheless, it is likely that the contact occurs within the lower argillaceous unit of Stith (1979), which is the equivalent of the “upper argillaceous member” of Calvert (1962).

Distribution:

As mentioned, Chazy Group strata are present in the subsurface of the Cincinnati Arch region, however the localized appearance of these units suggests the Cincinnati Arch was intermittently exposed and laterally interrupted by topographic relief. This relief was evidently related to lowered sea-level conditions associated with the Knox Unconformity, but may also indicate that the arch region was impacted by localized tectonic activity. Nonetheless, St. Peter and Wells Creek equivalents are recognized on both flanks of the Cincinnati Arch, and to the south of the Jessamine Dome in the Nashville Dome region where the equivalents include the Murfreesboro through the thin-bedded argillaceous Pierce Limestone. Eastward, the St. Peter – Wells Creek interval was correlated by Ryder (1991) into the Loysburg of Pennsylvania

Chazy Formations: St. Peter

On the western flanks of the Cincinnati Arch in the southern Illinois Basin, Nelson (1996 & 1998) described the St. Peter Sandstone as an interval of very pure (approaching 100 percent) fine-to-medium-grained quartz sandstone varying from white to light-brown in color. Quartz grains are often frosted (wind-blown?) and may display cross-beds and ripple marks. South and eastward the St. Peter thins substantially and is often reported in cores as absent. In central and eastern Kentucky, the St. Peter is a fine- to coarse-grained dolomitic sandstone which appears to be equivalent to the St. Peter to the west – but is somewhat more carbonate rich.

In the Cincinnati Arch region, the St. Peter is reported to reach a maximum thickness of 26 meters in southeastern Indiana in the vicinity of the Sebree Trough. The unit thins both westward and eastward into northern Kentucky and southwestern Ohio. In the central and eastern Jessamine Dome, the St. Peter is only about three meters thick. The St. Peter is located intermittently in the subsurface of Boone, Gallatin, Carroll and Henry Counties but ranges between zero and only two meters thick. It is not recognizable along the axis of the Cincinnati Arch, in Kenton, Grant, Owen and Harrison Counties although it is recognized in northern Pendleton County, Kentucky and Clermont County in southern Ohio on the east side of the arch (Carpenter, 1965). Throughout this region, the St. Peter appears to have some carbonate matrix although as it is a thin interval it is usually grouped with the lower Wells Creek.

Chazy Formations: Wells Creek & the Lower Argillaceous Unit

The "Wells Creek Dolomite" of Lusk (1927) was originally proposed as a relatively thick interval (over 100 meters-thick) immediately above the Beekmantown of the Wells Creek Basin in central Tennessee. In Tennessee, Bentall and Collins (1945) recognized the Wells Creek Dolomite as a 60 meter thick interval dominated by green, argillaceous dolostones and tan to green sandy dolomites. In the region of the Cumberland Saddle, in southern Kentucky, these same authors included up to about 130 meters of strata similar to this unit. Dever (1980) recognized the Wells Creek as a silty and dolomitic interval below the lowest micro-grained lithographic limestone (of the High Bridge Group) and Gooding (1992) suggested the top of the unit occurred where the interbedded dark gray shales of the lower High Bridge Group change to the dolomitic limestones of Wells Creek with occasional green shales. Thus it appears that the extent of the Wells Creek is still debated and not applied consistently.

To the north in northern Kentucky and southwestern Ohio, the lower Wells Creek, or what was referred to by Calvert (1962 a) as the lower dolomite member of the Chazy, is approximately 15 meters thick. It is characterized by a dolomite interval with a conglomeratic base that often contains thin lenses of sandstone overlain by finer grained, greenish-gray dolomite and at least one K-bentonite is known from this interval (Kolata et al., 1996). The unit commonly contains white, angular chert pebbles and is marked by thin interbeds of green and gray shale. Above the lower portion of the Wells Creek, the rest of the lower dolomite member is characterized by beds of light grayish-brown, finely crystalline, argillaceous dolomite. Near the top of the Wells Creek, the unit once again becomes interbedded with thin beds of green shale and a few argillaceous limestones similar to the Glenwood Shale of the Illinois Basin (Carpenter, 1965). This interval was recognized as the "lower argillaceous unit" by Stith (1979). It grades upward into more massive limestones in the transition to the Carntown unit of the Camp Nelson. These shales were also recognized in central Kentucky by Gooding (1992). As reported by Carpenter (1965) the lower member of Calvert's Chazy Limestone thus ranges in thickness from extinction to over 32 meters in Owen County, Kentucky. Although inclusive of the lower argillaceous interval of later workers, isopach maps show depo-centers in central Owen, and northern Grant counties in Kentucky, as well as in western Hamilton County in Ohio. A number of relatively small areas are noted where very little if any Wells Creek was deposited including in western Boone and western Grant counties in Kentucky, and southeastern Hamilton County in Ohio (Carpenter, 1965). The former sites lie west of the arch axis and immediately east of the inferred position of the Sebree Trough. The pattern suggests that the Cincinnati Arch was not a linear uplift feature at this time, but the region was likely a segmented series of small localized uplifts with intervening sub-basins perhaps resulting from relict topography of the Knox Unconformity.

Further north in the vicinity of the Findlay Arch of Ohio, Wickstrom and colleagues (1992) indicate that the Wells Creek in northwestern Ohio also ranges from 0 to as much as 18 meters-thick. It commonly averages between five and six meters, but thickens to the east. As in the south, it often contains dolomitic green shales, argillaceous limestones and dolomites. Wickstrom and colleagues also note the presence of some thin brown, gray and black shales interbedded with thin sandstones and siltstones. As in the southern portion of the Cincinnati Arch, the top contact was suggested to coincide with the lithologic change from the argillaceous carbonates of the Wells Creek to the relatively clean micritic limestone or dolomite of the Black River. Thus it is apparent that Wickstrom and others (1992) continue to use the concept of Carpenter (1965) for placement of the CBR contact at the base of what Stith (1979) referred to as the Carntown unit.

High Bridge Group: General Description, Contacts & Distribution

Outcrops in central Kentucky near the Kentucky River expose the High Bridge Group. The unit is characterized by a massive succession of shallow sub-tidal through supratidal micritic limestones, dolomitic limestones and dolomites. These were deposited during the larger scale transgressive-regressive cycle of the Blountian Tectophase of the Taconic Orogeny (Ettensohn & Kuhnhenn, 2002). The three main formations of the High Bridge include the Camp Nelson, Oregon, and Tyrone (Miller, 1905). The Camp Nelson is the thickest, most extensive, and most heterolithic unit of the High Bridge and contains a number of diverse lithofacies in comparison to the overlying Oregon and Tyrone formations. Collectively, these formations were deposited in tidal flat to protected shelf or lagoonal environments (Ettensohn et al., 2002c) and have been estimated to approach thicknesses of up to 215 meters (Wahlman, 1992). Average thicknesses are commonly about 125 meters. In the subsurface, Stith (1979) also recognized a tripartite Black River/High Bridge Group interval which he referred to as “lower third,” “middle third,” and “upper third” respectively. The lower third was in turn divided into three sub-units, which included the “lower argillaceous unit,” the “Carntown unit,” and the “upper argillaceous unit” (**see Figure 6**). The middle third was also defined and included two marker beds (I and II), and the upper third, with its large number of K-bentonites was defined to include at least six different marker beds (**Figure 7**).

In central Kentucky, although the Tyrone and Oregon have been fairly well-characterized (Cressman & Noger, 1976; Horrell, 1981; Kuhnhenn et al. 1981) the Camp Nelson has itself not been studied in such detail. Although it has some lithologic variability, to date it has not been sub-divided formally. Nonetheless, Ettensohn (1992) identified a few marker horizons that are useful in mapping at least locally and are most useful for correlation of weathered sections and have not been extended into the subsurface. These marker horizons include the lower white marker bed, the open-marine shale marker, the middle white marker, the mudstone marker and the upper white marker bed (in the Oregon Formation). Based on lithologic descriptions, relative thicknesses, and correlation of these marker beds the lower third and middle third of Stith (1979) are herein considered the equivalents of the Camp Nelson while the upper third is the equivalent of the Oregon and Tyrone formations (see **Figure 7**). In addition to correlation of the Pencil Cave K-bentonite of Ettensohn (1992) with the alpha marker bed of Stith (1979), other key correlations include recognition of the prominent calcareous green silty mudstone unit (marker Bed I of Stith) below the level of the Oregon Formation, recognition of the extensively bioturbated interval (marker bed II), and the correlation of the upper argillaceous unit of Stith

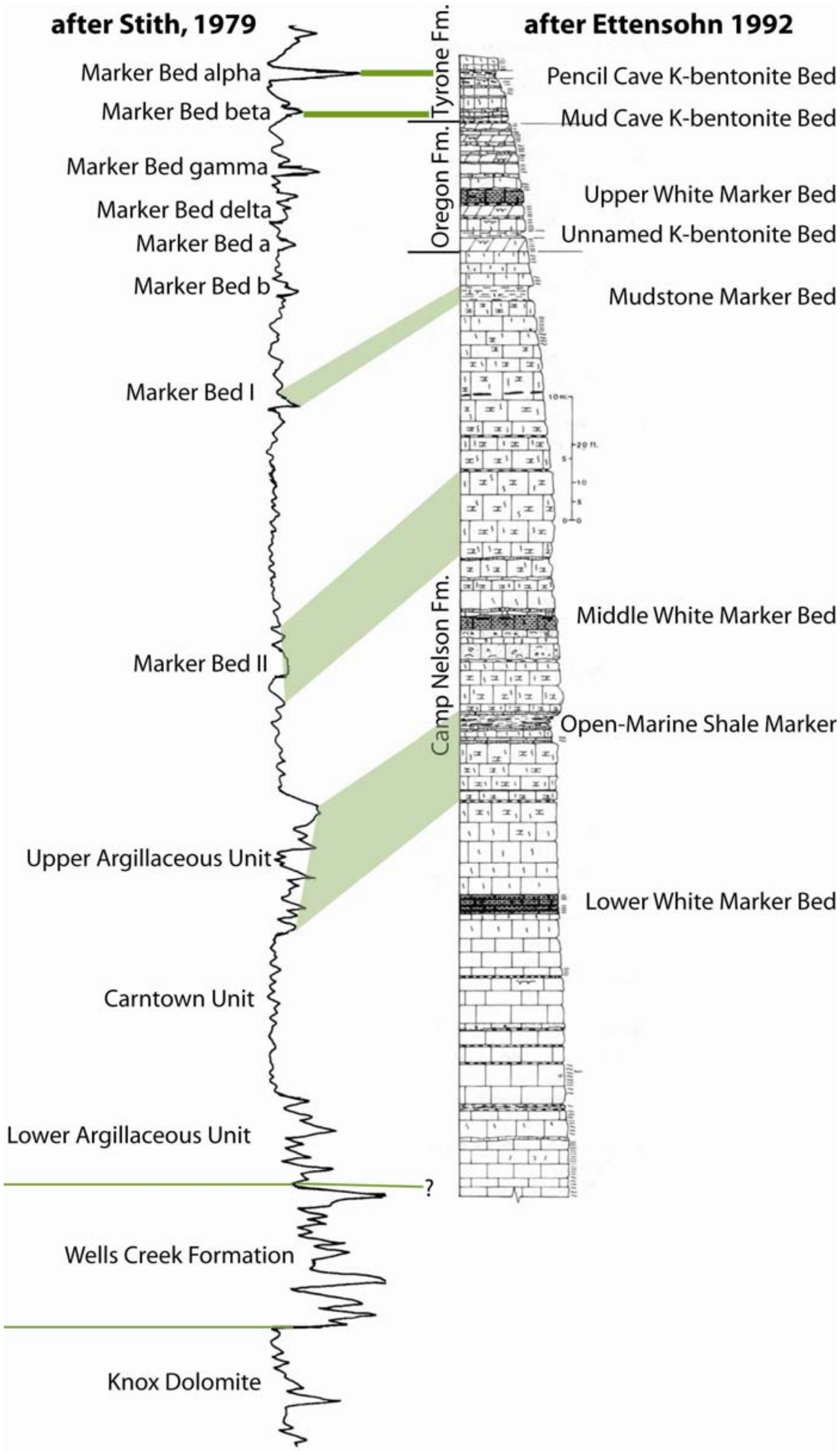


Figure 7: Stratigraphic marker horizons used by Stith (1979) in the subsurface of northern Kentucky (Gamma Ray Curve for Wilson #1 Well, Campbell Cty) correlated to those used by Ettensohn (1992) for outcrop exposures in the Camp Nelson region of the Kentucky River. As suggested by Kolata et al., 1996, the Pencil Cave of Ettensohn was correlated with the alpha K-bentonite of Stith and equated to the Millbrig K-bentonite.

with the open-marine shale marker of Ettensohn.

Black River Group Contacts

In Kentucky, the base of the Black River Group has been thought to coincide with the base of the Camp Nelson Formation. Unfortunately, the base of the Camp Nelson is not exposed in the Jessamine Dome and is only observable in the subsurface. As mentioned previously, the High Bridge Group is defined to overlie the Wells Creek Formation. However, details of a specific contact between these units are still in debate. In the subsurface of Ohio and northern Kentucky, both Carpenter (1965) and Stith (1979) considered the Wells Creek (lower dolomite member of Carpenter) to be equivalent to the Chazy Group. Nonetheless, Carpenter also included the "argillaceous unit" near the top of the Wells Creek within his Chazy suggesting the siliciclastic signatures (and gamma-ray response) were more like the Wells Creek. In contrast, Stith suggested the interbedded micritic limestones of the "lower argillaceous unit" to be most similar to Black River lithology and thus included the unit within the overlying Black River/High Bridge Group succession – a position also supported in the interpretations of Gooding (1992).

The same difficulty is recognized in northeastern New York State where the Black River is in contact with the underlying Chazy. In this latter region, the contact is generally placed at the transition from the fossiliferous medium-to-coarse-grained calcarenites (above an interval of interbedded argillaceous limestones) into the more massive buff brown-weathering dolomites (of the Pamelia Formation) which are in turn overlain by the typical Lowville lithologies (Oxley & Kay, 1959). In the Jessamine Dome, the lower argillaceous interval as recognized by Stith (1979) and Gooding (1992) may indeed be laterally equivalent of the fossil-rich argillaceous interval in the Valcour Formation of the Chazy. Therefore the overlying Carntown unit may at its base contain the true Chazy-Black River contact. Thus the Carpenter (1965) model is likely more analogous of the lithologic transition seen in the type region and the base of the Black River is drawn at the base of the Carntown unit of Stith.

The upper contact of the Black River equivalent (the High Bridge Group) has been placed historically at the top of the Tyrone Formation at the sharp erosional base of the coarse-grained Curdsville Limestone. The transition from cyclically-bedded peritidal micrites (Lowville or birdseye-type) into coarse-grained fossiliferous units of the basal Lexington is generally considered to be the Black River-Trenton contact (Ettensohn & Kuhnhenn, 2002. Historically, this contact has figured prominently in a number of different stratigraphic interpretations (see Figure 4). Nonetheless, recognition of this contact as the Black River – Trenton unconformity was predicated on the view that the Curdsville limestone contained a Kirkfieldian fauna (Twenhofel et al., 1954). Although the coarse-grained limestones of the Curdsville are lithologically distinct from the underlying Tyrone, the lack of appreciable siliciclastic interbeds in the Curdsville presents an argument against its classification as Trenton.

In the type region, one of the primary characteristics of Trenton lithologies includes the distinct interbedded nature of fossiliferous carbonates interbedded with thin dark, argillaceous calcisiltites and shales. In the Jessamine Dome and in the subsurface through northern Kentucky, the Curdsville is dominated by well-winnowed coarse-grained calcarenite and coquinal brachiopod rudstone facies with minor bedding plane partings and interbedded K-bentonite horizons. This facies is not a typical lower Trenton facies of the type region. In Kentucky, the first appreciable siliciclastic shale appears in the succession at the base of the Logana and approximately at the onset of the GICE. In New York State, the loss of shallow water birdseye micrites occurs well-below the top of the Black River Group although relatively pure deepening-upward, high-energy carbonates (Watertown through Selby Limestones) continue to be deposited up through the onset of siliciclastic deposition and the appearance of the GICE. Thus in Kentucky, depositional characteristics and the lithostratigraphic position of the Curdsville are remarkably similar to the type-region – albeit the Curdsville is arguably more coarsely-grained than the Watertown. This assessment was obviously shared by Nickles (1905) who felt that at least a portion of the basal Lexington was equivalent to the original concept of the Black River Limestone (Watertown-Selby of today **– see figure 4**). Nonetheless, herein, it is suggested that the top of the Black River Group be moved upward in the section from the base of the Curdsville to its top. This would position the Black River – Trenton contact in Kentucky at the analogous and likely synchronous position it occupies at the base of the Napanee Limestone (top Selby Limestone) in New York. However, in keeping with stratigraphic nomenclature as applied in Kentucky, the discussion herein retains the classification of earlier workers.

Distribution

The High Bridge Group (Black River equivalent) is exceptionally widespread in the subsurface of the Cincinnati Arch region. As reported by Cressman and Noger (1976) these supratidal, intertidal, and shallow subtidal carbonates were deposited over much of the GACB. These strata, the oldest units exposed in Kentucky, have been correlated directly into strata to the south in the Nashville Dome where the equivalent is referred to as the Stone’s River Group. The Stone’s River includes the Murfreesboro, Pierce, Ridley, Lebanon, and Carters Limestones (Wahlman, 1992); the latter three are likely equivalents of the High Bridge Group as exposed in the Jessamine Dome. In southern Indiana just west of the Jessamine Dome, the High Bridge grades laterally into units referred to as the Black River Group (Droste et al., 1982). As there are no exposures of the High Bridge Group equivalents in Indiana, stratigraphic nomenclature is based on correlation with the Black River Group of northern Ohio and correlation with units from the Illinois Basin. Thus the High Bridge is thought of as an equivalent of the Plattin Limestone and at least a portion of the Decorah Formation (Droste et al., 1982). To the east of the Jessamine Dome in the Appalachian Basin, the High Bridge is correlated into the Hurricane Bridge, Woodway, Ben Hur, Hardy Creek and Eggleston formations (Kolata et al., 1996). Northward, into northern Ohio and southern Michigan, the High Bridge is correlated in the subsurface to units referred to as Black River Group (Ryder, 1991, 1992).

Black River Formations: Carntown/Pecatonica

The Carntown is an informally recognized lithostratigraphic unit at the base of the Black River/High Bridge Group (after Stith, 1979) in the Cincinnati Arch region. The unit takes its name from a locality in Pendleton County, Kentucky on the Ohio River (Carntown) where the high-calcium limestone is mined in the subsurface however it is most often used as a stratigraphic term in Ohio. In this region on the eastern side of the Cincinnati Arch, the Carntown is brought closer to the surface near the core of a structure that is referred to as the Moscow-Carntown anticline. As suggested by Stith (1979) the Carntown unit is a distinct, pure limestone unit (upwards of 98 percent $CaCO_3$, with only minor dolomitic limestone), that occurs above the lower argillaceous unit and below the "upper argillaceous unit."

To the west of the Rough Creek Graben in western Kentucky, southern Illinois and southern Indiana a similar succession has also been recognized and has been correlated with the basal Platteville Formation (Pecatonica Member) of the Illinois Basin (Schwalb, 1969; Kolata and Noger, 1991). In the central Jessamine Dome a relatively pure limestone interval has also been recognized sandwiched between two argillaceous intervals (Dever, 1980). The pure limestone interval (up to 95 percent $CaCO_3$) was suggested as a possible equivalent of the Carntown (Noger and Drahovzal, 2005). Thus, Noger and Drahovzal correlated the high-calcium interval with the Carntown and consider it an equivalent of the Pecatonica of the Illinois Basin. These authors favor recognition of the pure limestone interval (i.e. the Pecatonica Limestone) as a new member of the Camp Nelson Formation of the High Bridge Group in place of the term Carntown. In their revised scheme, the High Bridge Group of the Jessamine Dome would be refined to include five lithostratigraphic units: 1) the "lower part of the Camp Nelson" (Stith's lower argillaceous unit), 2) the Pecatonica Member of the Camp Nelson, 3) the "upper part of the Camp Nelson" (Stith's upper argillaceous unit up to the level of his mudstone marker), 4) the Oregon Formation, and 5) the Tyrone Formation. Herein, it is believed that the lower argillaceous interval should be included in the underlying Wells Creek Formation and excluded as a member of the overlying High Bridge Group. Recognition of the Carntown/Pecatonica interval as a separate lithostratigraphic interval is necessary and supported herein. Nonetheless as this unit is only partially exposed in the Jessamine Dome, this interval might be considered as a separate formation of the High Bridge Group below the level exposed at Camp Nelson. Throughout this document, the use of Carntown is preferred over Pecatonica as it has been used in the stratigraphic literature of this region whereas Pecatonica has not.

Throughout the Jessamine Dome region, the Carntown is usually present in the base of the Black River. However it has been reported as being very thin to absent in a few cores due to onlap of existing topographic highs produced during the development of the Knox Unconformity (Stith, 1979). Thus the Carntown ranges from not present to upwards of about 22 meters thick.

Stith (1979) suggested that the unit is most commonly about 15 meters thick. Using wireline logs and the Jackson Township core (Brown County, Ohio), Stith (1979) recognized his Carntown to be a two part unit that ranged up to nearly 21 meters combined. Immediately above the lower argillaceous unit, the lower Carntown is composed of very pure peloidal micrites with minor dolomite areas (~12 meters). The upper Carntown is an interval of interbedded dolomites and dolomitic limestones (9 meters) immediately below the upper argillaceous interval. Stith (1979) interprets the Carntown as a subtidal through supratidal succession overall. At the Carntown lime plant, approximately 10 meters of the Carntown were mined for high calcium lime (Wolfe, 2008). This is similar to the thickness in Ohio just to the northwest of the Carntown mine. In the central Jessamine Dome in Fayette County Dever (1980) reports an interval of almost 17 meters to be of similar high purity. Unfortunately, very few fossils have been described from the Carntown interval although some corals including *Tetradium* and a larger tabulate – similar to *Foerstephyllum* have been noted in cores within the lower Camp Nelson.

Black River/High Bridge Group: Camp Nelson Formation

The Camp Nelson Formation of the High Bridge Group was named for over 87 meters of limestones incompletely exposed in the vicinity of Camp Nelson, Jessamine County, Kentucky (Miller, 1905). Miller recognized the Camp Nelson as a massive, conchoidal fracturing lithographic limestone. In the type area, Wolcott and colleagues (1972) and Cressman (1973) described the Camp Nelson as dominated by calcilutite facies with minor dolomitized and bioturbated intervals. Miller (1905) also noted the appearance of a few beds of white-weathering shale that ranged up to about a meter thick. Similar to the lower part of the Carntown, the Camp Nelson is dominantly a light dove-brown peloidal mudstone or calcilutite that often contains burrow mottling and a greater insoluble residue fraction than the Carntown (**Figure 8**). Kuhnhenn and colleagues (1981) and Ettensohn and colleagues (2002d) interpret the Camp Nelson to be dominated by shallow-marine limestones with occasional interbeds of peritidal limestones and dolomitic limestones. In addition, the succession is interrupted by several prominent marker horizons. In weathered outcrops, Ettensohn and colleagues (2002d) recognize: two white calcilutite beds (both are laminated, mud-cracked, birdseye micrites), a prominent fossiliferous open-marine shale, and a white-weathering dolomitic mudstone near the top of the Camp Nelson Formation (**see figure 7**). The shale-dominated units are likely the shale units recognized earlier by Miller (1905).

Although they have not been used for correlation, the Camp Nelson often demonstrates small, meter-scale cycles which have been compared to the punctuated aggradational cycles (PAC's) of Goodwin and colleagues (1985; 1986). Camp Nelson cycles often contain a thin basal calcarenite to intraclastic breccia sitting above an erosional base that may be somewhat channelized (Kuhnhenn et al., 1981). This is in turn overlain by successions of massive often bioturbated peloidal calcilutites that grade upward into ribbon-bedded calcilutites and dolomitic limestones which are often capped by laminated dolostones. The peloidal micrites that are so dominant in the Camp Nelson are the predominant facies in most of these cycles throughout the central Jessamine Dome and along the Cincinnati Arch to the north.

In contrast, portions of the Camp Nelson become more coarse-grained and fossiliferous to the east of the central Jessamine Dome on the eastern margin of the Kentucky River fault zone near Boonesborough (**figure 9**). Core KGS-CK-15 (Mary Brandenburg Farm near Winchester, KY housed at the Kentucky Geological Survey Core Repository), shows many of the marker horizons identified in the Jessamine County region. However, the cyclically-bedded peloidal

Figure 8: Representative facies of the Camp Nelson Formation of the High Bridge Group from cores from central Kentucky and southwestern Ohio.

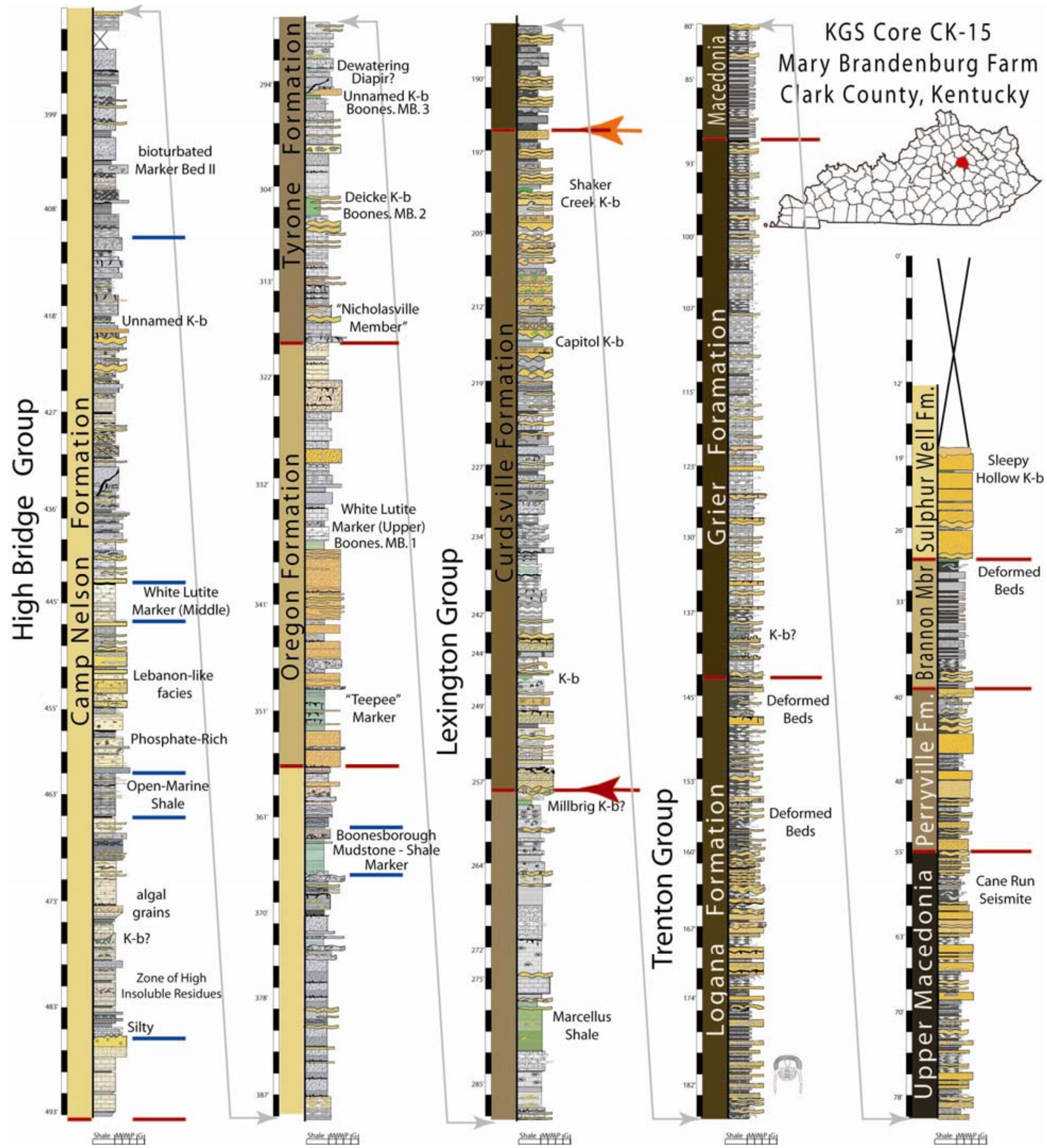


Figure 9: Stratigraphic section for the upper Camp Nelson through lower Lexington Limestones from a core from Clark County, Kentucky (KGS Core CK-15) located on the eastern edge of the Cincinnati Arch along the southern extension of the Moscow-Carntown anticline.

calcilutites are typically developed into interbedded, fine-to-medium-grained wackestones and grainstones that contain a larger diversity fauna than that occurring in the outcrops to the west. Rather than the typical peloidal micrites of the Camp Nelson, in the short-distance from Jessamine to Clark County (~45 km), this interval becomes reminiscent of the facies classified as the Lebanon Limestone in central and eastern Tennessee to the south and east. Thus, eastward across the Jessamine Dome, the Camp Nelson shows a transition into somewhat coarser-grained (although still micritic) facies than are exhibited in the outcrop region. This facies transition may be related to syndepositional fault movement during the Blountian phase of the Taconic Orogeny.

Faunally, the Camp Nelson is apparently more fossiliferous than the underlying Carntown unit although in the type area fossils are still fairly restricted except for in a few horizons. Nosow & McFarlan (1960) reported the Camp Nelson to be dominated by a bryozoan-mollusk fauna. These authors recognized the gastropod *Maclurites bigsbyi*, and a number of cephalopods including *Cryptophragmus* which enabled its correlation with the Pamelia and Lowville formations of New York. Kuhnhenn and colleagues (1981) and Ettensohn and colleagues (2002 d) recognized additional faunas in the unit indicating that most beds can be characterized by a sparse, molluscan-ostracod-tabulate coral fauna. However, in the bases of the shallowing-upward cycles, more open-marine forms are characteristic. These include *M. bigsbyi*, and orthoconic and planispiral cephalopods. In the open-marine shale intervals additional forms have been recognized and include brachiopods, bryozoans, gastropods, crinoids, sponges, corals (rugose and tabulate), ostracods, trilobites, and numerous algae and occasionally stromatolites (Ettensohn et al, 2002 d). Another important texture of the Camp Nelson is the occurrence of bioturbation especially near the top of the formation which suggests infaunal organisms were also important components of the assemblage. Many of these taxa and sedimentary textures become more prevalent in the CK-15 core where corals and echinoderms become more predominant components of the faunas, although other organisms are certainly recognized.

Given the range of lithologies and faunas in the Camp Nelson, Ettensohn and colleagues (2002 d) suggested the Camp Nelson was deposited in a partially restricted, shallow-ramp, lagoonal environment similar in construct to the low-energy regions of the Bahamian platform. In the subsurface further north, Stith (1979) recognized the majority of the Camp Nelson equivalent (below his Marker bed I – mudstone marker) to be dominated by subtidal facies with only occasional intertidal to supratidal cycle caps. The interval is also noted to show increased siliciclastics and occasional carbonaceous shales, and the micrite-rich intervals are darker gray in color (as opposed to dove) and have somewhat lowered occurrence of magnesium-rich dolomitic limestones suggesting that the amount of restriction was somewhat less north of the Jessamine Dome. Stith (1979) also reports the lowest occurrence of a K-bentonite just below the level of his bioturbated marker bed (marker bed II). A K-bentonite is also located in a similar position in the CK-15 core (**see figure 9**). The combined evidence for 1) the increased mineralization (phosphatization and iron-stained hardgrounds), 2) insoluble residues, 3) detrital silts and clays, and 4) the first occurrence of the K-bentonites in the Camp Nelson, suggests that the Blountian Tectophase was not only well underway, but that tectonically-derived sediments were indeed delivered in increasing quantities to the region as the feather-edge of the Sevier basin clastic fill. Moreover, it is apparent that local fault activity along the Cincinnati Arch may have contributed to lateral lithologic change. Thus small-scale cyclicity and subsequent shallowing of facies near the top of the Camp Nelson was likely enhanced by both sea-level change and localized tectonic activity associated with the Blountian Tectophase. Ettensohn (1992) suggested the widespread nature of the tidal-flat lithologies of the later Camp Nelson may have been related to a period of Blountian flexural relaxation.

Black River/High Bridge Group: Oregon Formation

The Oregon Formation, commonly referred to as the Kentucky River Marble (Nosow & McFarlan, 1960), has been recognized as a distinct stratigraphic unit of the High Bridge Group. Named after localities on the Kentucky River in Mercer County, the Oregon was originally classified as an interval of between 7 and 9 meters of buff, yellowish gray, even bedded "highly magnesian limestones," which are especially prevalent in the lower and upper portions of the unit (Miller, 1905; **figure 10**). Wahlman (1992) defined the Oregon as a unit composed of finely crystalline dolomite interbedded with micritic limestones. Most workers recognize the contacts of the Oregon from the base of the lowest dolostone through the top of the highest dolostone bed (Cressman & Noger, 1976). The Oregon and overlying Tyrone have received appreciably more study than the underlying Camp Nelson (Cressman & Noger, 1976; Horrell, 1981; Kuhnhenn et al., 1981; Conkin & Conkin, 1983; Ettensohn et al., 2002 c, d etc.).

Ettensohn and colleagues (2002 d) indicate that the Oregon is an arbitrary unit based on the occurrence of mappable dolostones above the Camp Nelson and below the Tyrone Formation. Across the Jessamine Dome, the Oregon thickens from west to east. In the southwestern portion of the Kentucky River outcrop area (southern Herrington Lake, Boyle County), the Oregon was measured at about 1.9 meters and to the east the unit thickens to over 19.8 meters at Boonesborough in the vicinity of the eastern Kentucky River fault zone in Clark County (Wolcott & Cressman, 1971; Black, 1968). In Anderson County south of Frankfort, KGS Core C-204 (Lawrenceburg area) shows just less than 15 meters of Oregon, although the upper portion of the Oregon is thinner and less distinctive. Further west and south of the Jessamine Dome, the Oregon is not recognized and is not correlated in the subsurface by Noger and Drahovzal (2005). To the northeast of the Jessamine Dome in the Clark County core only about 12.2 meters of Oregon are developed suggesting thinning to the east in the vicinity of the

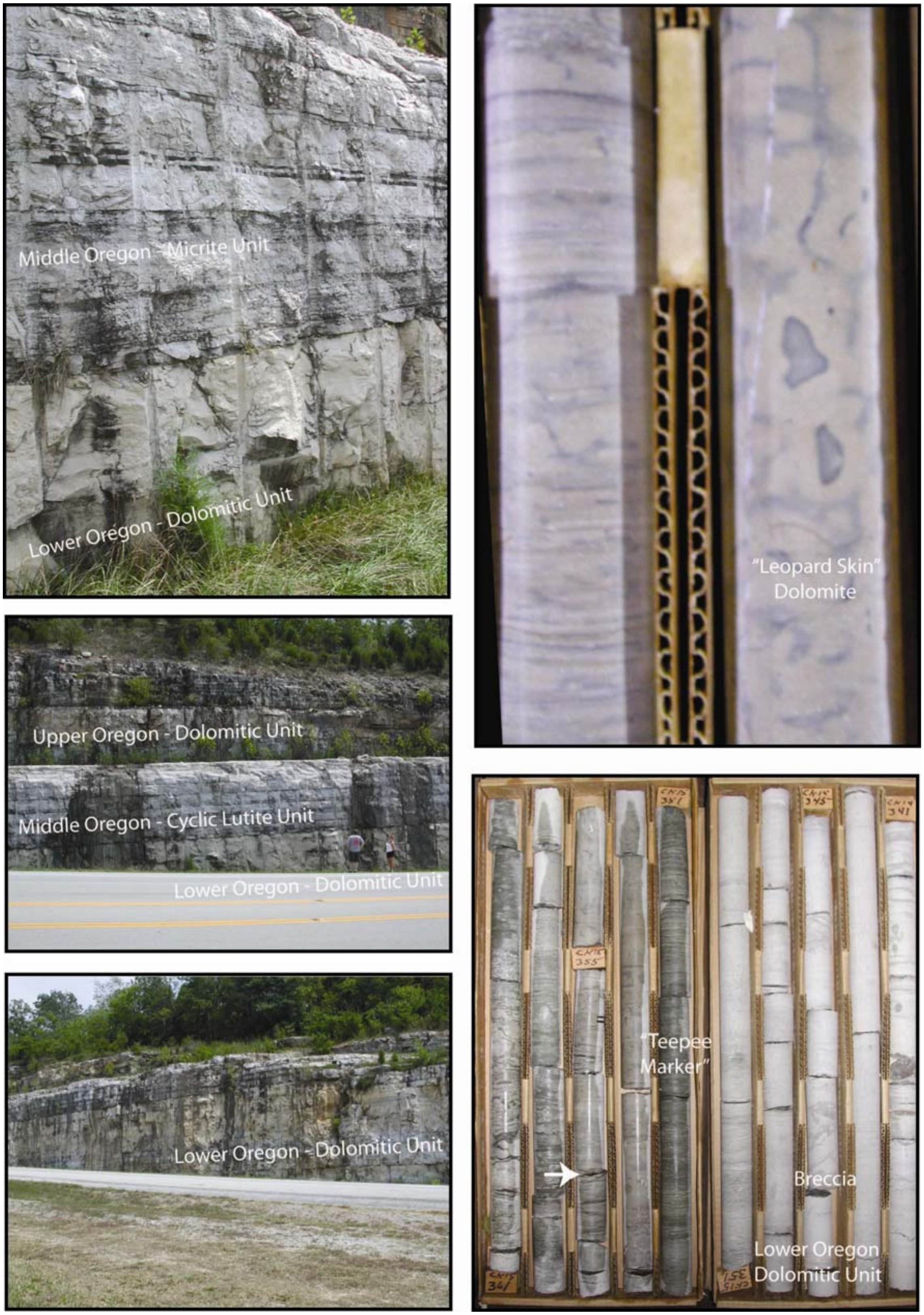


Figure 10: Outcrop and representative core photographs of the Oregon Formation of the High Bridge Group. The outcrop photographs are from KY Rte 627 just north of the Memorial Bridge over the Kentucky River at Boonesborough. Core photographs are from the Mary Brandenburg Farm Core KGS-CK-15, in Clark County, Kentucky located just northeast of Boonesborough.

Rome Trough. Across this region (~ 50 km), the base of the Oregon appears to be relatively synchronous throughout its outcrop area as suggested by the position of the first dolomite bed relative to the upper mudstone marker near the top of the Camp Nelson (Cressman & Noger, 1976). Thus the eastward thickening of the formation results from the recognition of additional dolomitic "leopard skin" beds higher in the section where the top of the Oregon is reported to inter-tongue with the base of the Tyrone (Kuhnhenn et al., 1981). Because the contacts (especially the upper contact) are hard to recognize, the unit may be a diagenetic unit which formed due to the presence of precursor lithofacies that were prone to secondary dolomitization in the vicinity of fault and fracture zones of the region (Black & Haney, 1975; Ettensohn et al., 2002d).

Nonetheless, there are a number of lithologic indicators that support a primary supratidal origin for at least some of the Oregon carbonates. The Oregon can be defined as a tripartite unit with a lower and upper dolomite-rich interval and an intervening micritic limestone or "lutite unit" (**see Figure 10**). Although not formally recognized as a sub-unit, the middle limestone unit is generally associated with the white lutite marker recognized by Ettensohn and colleagues (2002 d). Although the middle beds are dominated by micritic limestones, they also contain stringers and lenses of coarser-grained fossiliferous beds (**figure 11**). The lower and upper Oregon sub-units are composed of laminated to bioturbated "leopard-skin" crystalline dolomites. Laminated units are often planar to crinkly laminated (suggesting microbial mat origins) and show alternations between coarser and finer dolomites. These beds are typically interbedded with layers of intraclastic breccias and mud-cracked intervals. As these beds also exhibit cyclicity, they are also interbedded with micritic limestones and greenish grey argillaceous seams that are similar to the mudstone marker in the top of the Camp Nelson.

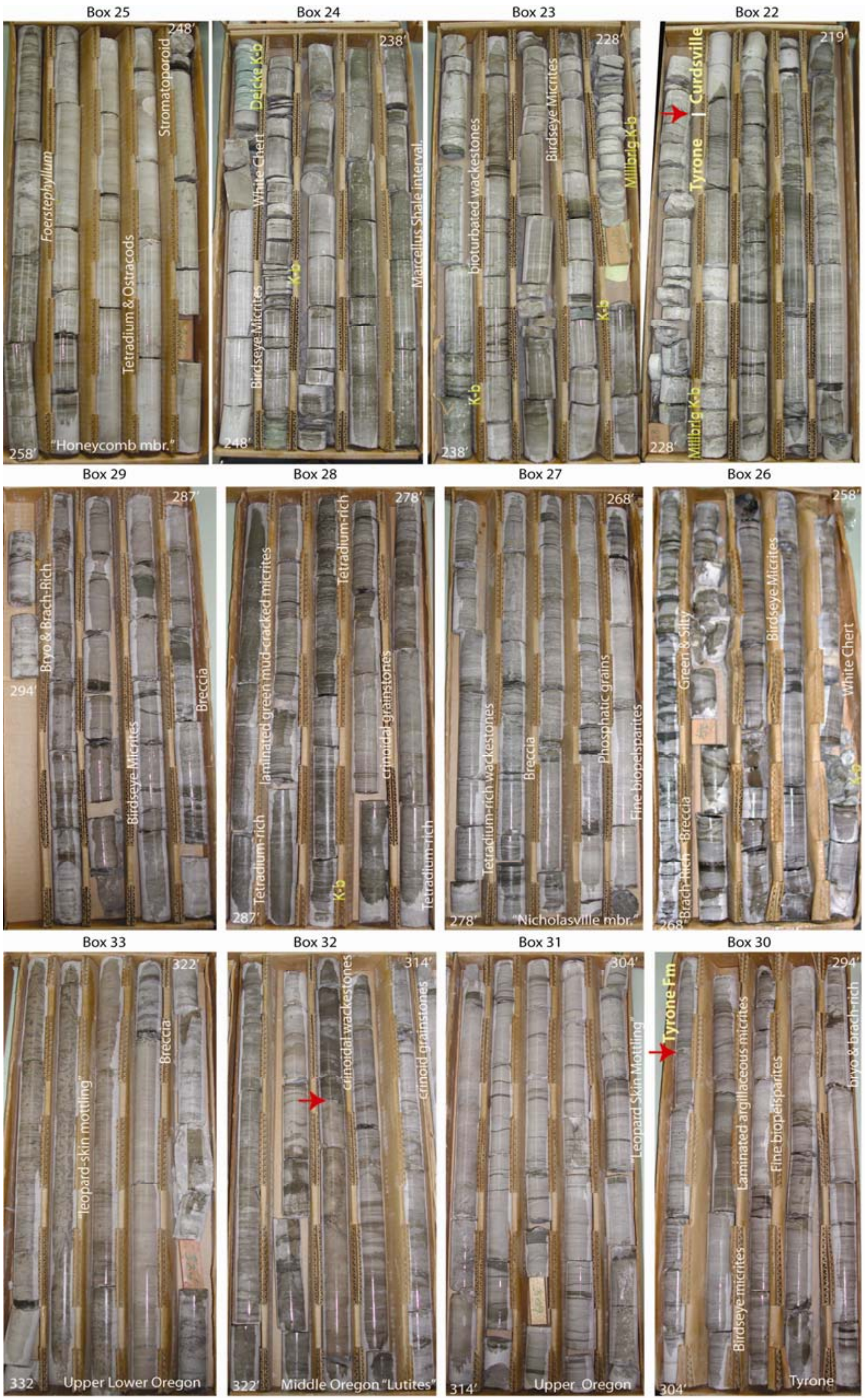


Figure 11: Photographs of KGS-C204, Lawrenceburg, Kentucky (Anderson County) core showing the Oregon through Curdsville interval. Portions of the upper lower Oregon, middle Oregon, Upper Oregon, and the entire Tyrone are displayed. Key K-bentonite horizons and other marker units are identified in the Tyrone as discussed in the text.

The propensity of this interval to exhibit a diversity of lithologies, while still preserving supratidal to subtidal depositional structures, original lamination, and a number of limestone facies, suggests that at least some of the Oregon is primary in origin. The loss of the upper Oregon supratidal facies to Tyrone-type facies west and east of the Boonesborough axis suggests perhaps that the Boonesborough region may have been slightly more elevated and was the last area to succumb to deepening or basin relaxation according to the Ettensohn (1992) model mentioned previously.

As noted, the Oregon, has not formally been broken into sub-units, and is usually considered to represent a mosaic of up to as many as seven different lithofacies (Horrell, 1981) deposited in shallow subtidal through intertidal and supratidal environments. Despite the heterogeneity of facies, Horrell (1981) reported that statistical analyses showed that these lithofacies are arranged into shoaling-upward (shallowing-upward?) cycles capped by laminated intervals. With recognition of an increasing number of internal marker horizons in the Oregon, it has been possible to correlate many of the cycles of the Oregon into time equivalent facies elsewhere on the Cincinnati Arch.

Marker horizons in the Oregon include: 1) a laminated green, argillaceous and evaporitic (?) dolomitic interval with possible teepee structures located just above the base of the Oregon (as per Kuhnhenn et al., 1981; but interpreted here as underlying the Oregon), 2) the white lutite beds (recognized by Ettensohn et al., 2002 c, d), and 3) a number of K-bentonite and possible K-bentonite horizons recognized by Conkin & Conkin (1983) and Kolata and colleagues (1996). Cressman and Noger (1976) recognized a prominent K-bentonite just below the last massive dolomite at Boonesborough and correlated the bed with a K-bentonite in the lower Tyrone at Camp Nelson approximately 16 meters below the Deicke K-bentonite (their Pencil Cave K-bentonite). This K-bentonite was correlated with Stith's (1979) b K-bentonite by Kolata and colleagues (1996) and to the Hockett K-bentonite of the Appalachian Basin. It is likely that this prominent K-bentonite is also the equivalent of the "first Boonesborough Metabentonite" (of Conkin & Conkin, 1982). This particular K-bentonite does not appear to be present in the Lawrenceburg core shown in figure 11. In addition to this latter K-bentonite, Conkin and Conkin (1983) recognized another 17 "metabentonite" seams at Boonesborough and considered the majority of them to occur above the Oregon (their interpretation of Oregon at this locality was constrained relative to previous workers). Kolata and colleagues (1996) consider the majority of these K-bentonites, owing to their relatively thin nature, to be of limited use in correlation. Nonetheless, both the prominent first Boonesborough Metabentonite, and the superjacent second Boonesborough Metabentonite are well-developed and are useful in correlation (**see figure 9**). Seven K-bentonites are well-developed in the Lawrenceburg core and exhibit swelling behavior when wetted. Other potential bentonitic horizons are recognized although these are bioturbated into the carbonate lithologies and are less distinct.

Although the dolomitized intervals in the Oregon are dominantly fossil poor, the coarse-grained breccia beds and pelsparites at the base of small-scale cycles especially in the middle unit are fossiliferous as in the underlying Camp Nelson. Unfortunately, no detailed faunal list is known for the interval. However, fossils recognized from the Oregon include a fauna dominated by ostracodes and *Tetradium* corals but also include crinoids, small twig bryozoans, occasional brachiopods, and minor trilobite fragments. In addition stromatolites of several morphologies have also been recognized as an important component of the Oregon fossil assemblages, as are numerous trace fossils including vertical burrowing and borings, as well as bioturbated fabrics that were enhanced by dolomitization.

Black River/High Bridge Group: Tyrone Formation

The uppermost formation of the High Bridge Group is the Tyrone Formation (Miller, 1905). The Tyrone was named for a locality of that name on the south bank of the Kentucky River in Anderson County in the western Jessamine Dome. In this region, exposures of relatively pure limestones and interbedded clay and clay-shale layers (ranging from a few cm to over one meter thick) are well developed and are usually well-exposed in the Kentucky River Gorge and in road cuts nearby. Similar to portions of the Camp Nelson, the Tyrone is composed of relatively light, dove-colored conchoidal fracturing limestones. The Tyrone is typically composed of sub-tidal biopelsparites, biomicrites, and peritidal micrites that are similar to some portions of the underlying Oregon Formation, except that they contain significantly less dolostone and are interbedded with argillaceous units (Wahlman, 1992; **see figure 11**).

Given that the contact with the underlying Oregon is defined at the position of the highest dolostone bed, the base of the Tyrone is somewhat arbitrary and subject to variable interpretation, but is generally regarded as a conformable contact. In the Clark County core, the base of the Tyrone is placed at the base of a prominent shallowing-upward cycle floored by an intraclastic bed that grades into an interval of peloidal micrites interbedded with fossiliferous ribbon grainstones. The intraclastic bed occurs about six meters above the first Boonesborough K-bentonite (of Conkin & Conkin, 1983) and about 3.6 meters below the level of the second Boonesborough K-bentonite recognized as the Deicke K-bentonite. Elsewhere this contact is not as well-developed. In the Lawrenceburg core, the contact has been drawn at the top of the highest prominent "leopard-skin" below the first major interval of peritidal micrites and fine-to-medium grained biopelsparites containing stringers of brachiopod and bryozoan-bearing wackestones. In this location, the contact is thus about 16 meters below the Deicke K-bentonite.

The lithologic transition out of the Oregon into the Tyrone represents renewed circulation as somewhat more offshore facies are deposited under slightly increased energies than are developed in the Oregon. Regionally, despite the challenge in locating a specific contact, the transition is similarly noted by most workers (Cressman & Noger, 1976; Kuhnhenn et al., 1981).

The upper contact of the Tyrone is one of the most recognizable contacts in the Jessamine Dome region, and has been discussed in previous studies. Ettensohn (1992) and Ettensohn and Kuhnhenn (2002) identified the contact as unconformable in the Camp Nelson area south of Nicholasville, Kentucky. In this area, the contact between the High Bridge Group and the Lexington Group is closely allied with the position of the Millbrig K-bentonite (Mud Cave of earlier workers; **figure 12**). The Millbrig has a variable thickness but is often immediately overlain by the coarse grainstones, coquinal rudstones, and brecciated lag beds in the base of the overlying Curdsville Limestone. In many instances, the contact can be observed to truncate the underlying Millbrig and completely remove the K-bentonite and underlying units within the span of single outcrops – as is apparent at Camp Nelson on KY Rte. 27, and in the Marcellus, Kentucky road-cuts on KY Rte. 34 near Herrington Lake. Elsewhere, the Millbrig has been completely truncated along with a couple of meters of additional underlying strata – as appears to be the case at Frankfort, Kentucky. Ettensohn and Kuhnhenn (2002) observed significant evidence for formation of lithoclasts (rip-ups) from the underlying Tyrone, as well as reworking and abrasion of corals and other faunal elements in the base of the Curdsville. In addition, these geologists note the occurrence of coarse-grained facies that may be piped down into the fine-grained lutites of the underlying Tyrone, which they suggest are Neptunian dikes. Alternatively, these coarse-grained pipes may in fact be let-downs into an irregular, mineralized, karstic unconformity. The contact is shown in figure 12 as exposed at the Marcellus, Kentucky outcrops

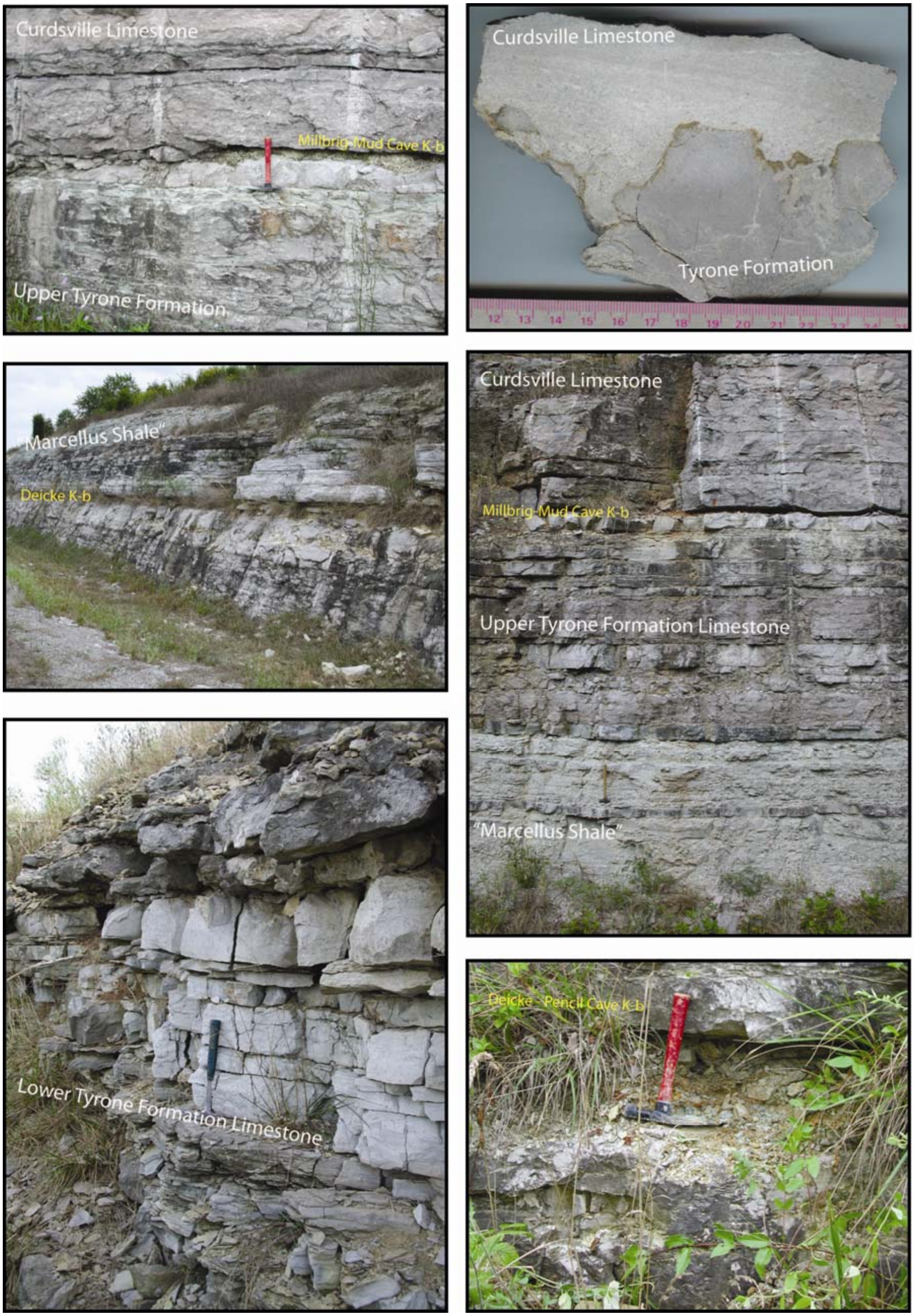


Figure 12: Outcrop photographs of the Upper High Bridge Group and its contact with the overlying Curdsville Limestone near Herrington Lake, Marcellus, Kentucky. Images show some key marker horizons found in the Tyrone Formation.

(upper left and center right) with the sectioned slab (upper right) coming from Frankfort, Kentucky (on KY Rte. 627). At Marcellus, the upper Tyrone is a greenish-silty argillaceous calcilutite that may itself contain bentonitic materials. This in turn is overlain by a 15 centimeter birdseye micrite capped-by the recessive- weathering soft clays of the Millbrig K-bentonite. In this locality, where the Millbrig is truncated along the outcrop, the coarse-grained Curdsville truncates the underlying micrites, but does not exhibit the same well-developed karstic surface as shown in figure 12 or as described by Ettensohn and Kuhnhenn (2002) at Camp Nelson.

Given these contacts, Cressman and Noger (1976) indicate that the Tyrone ranges from 17 to 47 meters thick. The thickest portion of the Tyrone is located on the western side of the Jessamine Dome where the Oregon is considered to be much thinner and the basal Tyrone contact extends down to near the top of the Camp Nelson. The unit shows substantial thinning to the eastern side of the Jessamine Dome including in the vicinity of Boonesborough. In the intervening area in Anderson County (Lawrenceburg Core), the Tyrone is 22 meters thick and in the Clark County core north east of Boonesborough (**figure 9**), the Tyrone is 18.9 meters thick (although the Millbrig is recognizable). Without recognition of the prominent Millbrig K-bentonite at Boonesborough, it appears that some thinning may be due to truncation or non-deposition of the Tyrone at different times in this area, but is also attributed to lowering of the Oregon-Tyrone contact by loss of the well-developed upper Oregon unit. In the subsurface to the north of the Jessamine Dome, Stith (1979) indicates that the Tyrone equivalent (his upper third of the Black River Group) is about 38 meters thick up to the level of his marker bed alpha (Millbrig K-bentonite; Kolata et al., 1996).

Lithologically, the Tyrone is the most birdseye-rich (fenestral) interval of the entire stratigraphic succession in Kentucky and was referred to as the “Birdseye Limestone” by early workers including Nickles (1905; **see figure 4**). Nosow and McFarlan (1960) indicate that the Tyrone is easily recognizable based on the occurrence of the prominent K-bentonites (Mud Cave near the top and Pencil Cave near the middle of the formation), as well as nodules, lenses, and even beds of chert usually located in the immediate proximity to the prominent K-bentonites. In addition, Stith (1979) recognized at least six marker beds, five of which are considered well-developed bentonites and bentonitic shales. Despite this fact, no formal lithostratigraphic division has been applied to these strata as carbonate lithologies are considered to be highly variable between most outcrops. However, Nosow & McFarlan (1960) recognized the "upper half" of the Tyrone bracketed by the Pencil Cave (Deicke K-bentonite) below and the Mud Cave (Millbrig) above and a "lower half" of the Tyrone (immediately below the Pencil Cave). These authors informally sub-divided the lower half into two units in the southwestern portion of the Jessamine Dome (Boyle, Garrard, Jessamine and southwestern Fayette counties).

At Camp Nelson, these authors recognize a seven-meter thick interval immediately below the Pencil Cave as a Camp Nelson-like bioturbated unit, which these authors referred to as the "honeycomb member." The bioturbation here is distinct in that it appears to have occurred as vertical and sub-vertical *Phytopsis*-style burrows produced in firm micritic sediment. These burrowing organisms did not produce the homogenized fabrics as recorded elsewhere in the section and were evidently domichnial rather than fodichnial traces. The burrows are often set off and recognized by darker brown, micrite infillings that can be dolomitic – but are generally not spar-filled. Cressman & Noger (1976) described this unit as a massive (~10 meters-thick) interval of micrite and biopelmicrites with dolomite-filled burrows.

Immediately underlying this unit Nosow and McFarlan (1960) recognized ribbon-bedded coarse-grained facies interbedded with fine micrites and argillaceous seams. This interval ranged up to three meters-thick in their analysis and was referred to as the "Nicholasville member" of the Tyrone Formation. The latter unit (not to be confused with the Nicholas Member of the Cynthiana Formation) only appeared to be well-developed in Jessamine County outcrops near Camp Nelson. This is the same unit recognized by Cressman and Noger (1976) to contain pelsparites and biopelsparites about five meters thick and located just above the top of the Oregon Formation in this area. Although not previously recognized outside the Jessamine County region, given the approximate position of the Nicholasville member relative to the Pencil Cave K-bentonite, the intraclastic breccia overlain by the thin grainstone stringers at the base of the Tyrone at Boonesborough maybe a lateral equivalent of this unit on the eastern margin of the Jessamine Dome: however, it is certainly less-well developed. In the Anderson County core (**figure 11**), a similar interval is recognized below the "honeycomb" facies and shows several shallowing-upward cycles, each with a coarse, often brecciated, base shallowing upward through fossiliferous (brachiopod, bryozoan, crinoid, coral-rich) wackestones into birdseye micrites and laminated calcilutites. In a few cases, cycle caps show faint dolomitic limestone with "leopard-skin" fabrics, but these do not appear to repeat above. Thus, like the Oregon, the Tyrone can be divided up into roughly three sub-units: the "Nicholasville member," the "honeycomb member," and the bentonite-rich "upper member." Collectively these units show a transition from open-marine shallow sub-tidal influenced facies (and even coarse-grained) lithologies upward into more restricted intertidal facies of typical upper Tyrone. Even with the upward shallowing and restriction, facies in the upper Tyrone rarely develop major dolomite units like the underlying Oregon.

In addition to the prominent bioturbated peloidal micrites, biopelsparites, and fenestral micrites mentioned previously, the Tyrone contains two prominent and relatively thick green calcareous shale units commonly interbedded with thin silty micrite beds. The lower of these units is developed in the lower portion of the Tyrone and is recognized as the delta marker bed of Stith (1979). The delta marker bed is described as a thick, very argillaceous limestone (up to 4.5 meters-thick), and as demonstrated in the Anderson County core is relatively green and silty in appearance. Cressman and Noger (1976) recognize a similar bed in the western Jessamine Dome. In this region it is located just below the level of their "Nicholasville member" equivalent and these authors describe the unit to be mud-cracked, but containing occasional marine fossils, such as: *Tetradium* corals, and gastropods. "Mud-chip conglomerates" also occur especially near its top. Eastward at Boonesborough, this same unit is not as well developed and it is not clear that it is present and no similar unit is exposed in the Clark County core (**see figure 9**).

The other prominent silty argillaceous calcilutite or micrite unit occurs above the level of the Pencil Cave below the level of the Mud Cave K-bentonites near the top of the Tyrone Formation (**see figures 11 & 12**). It is well-developed and exposed in the western Jessamine Dome region where it has been informally referred to as the "Marcellus shale" (Brett et al., 2004). As this name is already used and has priority in the Devonian, a better name might be Herrington Lake shale. Nonetheless, this prominent green-shaly unit in the top of the Tyrone is exposed in a west-to-east transect from just east of the Kentucky River at Marcellus, Kentucky through the Camp Nelson type area, and into the Boonesborough area where it is prominent above the Deicke equivalent bed. In the former region, the core from Lawrenceburg (KGS C-104) shows a well-developed "Herrington Lake" interval that ranges up to about 2.5 meters and is clearly recognized below the Millbrig and at least two other K-bentonites. At Boonesborough a satisfactory Mud Cave K-bentonite has not been recognized as it appears to be truncated at the position of the Tyrone-Curdsville contact. Nonetheless, the occurrence of the Herrington Lake shale, a short distance below the unconformable contact, helps to calibrate the amount of erosion that occurred at the contact. Just northeast of Boonesborough in the Clark County core, a poorly developed greenish laminated micrite interval is also recognized, but in this locality it again is preserved below the Mud Cave (Millbrig) K-bentonite (**see figure 9**) suggesting that less truncation at the top of the Tyrone occurred in this region.

In contrast, to the north including in exposures at Frankfort, Kentucky only the lower shale-rich interval is developed below the major K-bentonites of the Tyrone. At Frankfort, it appears that both the Millbrig and the Herrington Lake shale are truncated at the base of the Curdsville, although the Pencil Cave (Deicke) and at least one additional K-bentonite above the Deicke are recognized. Farther north in the subsurface, cores show that although the Millbrig is again recognized below the top Tyrone unconformity, but the Herrington Lake shale interval is not distinct and a nodular bioturbated wackestone facies is developed in its place in cores from the Cincinnati region. Combined with the fact that the Millbrig and Deicke (Stith's marker beds alpha and beta respectively) become more closely spaced, the loss of the silty argillaceous sediments in this region of the Jessamine Dome suggests that the main depocenter for the upper High Bridge Group was centered in the southern portion of the Jessamine Dome. Moreover, it is also likely that the siliciclastic component of these argillaceous beds was a progradational product of transport from a more southern and southeastern source area. This was likely the Blountian highlands in the southern Appalachians. The loss of the Millbrig and underlying Herrington Lake shale in areas further south was likely related to movements along ancestral Kentucky River faults that became periodically reactivated during the later phases of the

Blountian Tectophase leading into the tectonic impact of the Vermontian Phase that activated during the early Trenton.

According to Wahlman (1992) and Frey (1995) fossils are sparse in the Tyrone, especially in the upper Tyrone. Unfortunately, very few taxonomic lists have been located for any of the exposed High Bridge Group units in Kentucky; most are limited to specialized studies (see below). Most fossil occurrences are thus limited to general listings rather than specific taxa that would be useful for major stratigraphic correlation. For instance, Cressman and Noger (1976) recognize a marine assemblage in the lower Tyrone. Below the level of the Pencil Cave, occasional beds are occasionally rich in brachiopods, gastropods, bryozoans, bivalves, ostracods, trilobites, and crinoids (i.e. in the "Nicholasville member" of the Tyrone as per Nosow & McFarlan; 1960). In addition, colonies and numerous fragments of the tabulate coral *Tetradium cellulosum* are also pronounced components of the Tyrone, as are numerous vertical to sub-vertical burrows. Fortuitously, some of the cores studied in this investigation show a number of important taxa including a species of stromatoporoid (likely *Stromatocerium rugosum*) as well as the tabulate coral *Foerstephyllum halli* and at least two different forms of *Tetradium*. However, while there are a few fossil-rich horizons in the lower Tyrone, the upper Tyrone is exceedingly depauperate with a much more restricted group including gastropods, ostracods, *Tetradium*, and a number of nautiloid cephalopods.

Ettensohn and Kuhnhenn (2002), indicate that both planispiral and orthoconic forms are represented in the upper Tyrone and are similar to forms found in the Turinian through Rocklandian of New York and Ontario. Frey (1995) did produce a comprehensive list of nautiloids from the lower Tyrone Formation collected from an outcrop just south of Sulfur Well in Jessamine County. A single collection was made from a level 8 meters below the Deicke

(Pencil Cave) K-bentonite from a coarse-grained biopelsparrudite, likely at the level of the Nicholasville member of Nosow and McFarlan (1960). Frey identified as many as 18 species, 14 genera, and 6 orders in this silicified assemblage. In addition to numerous other fossils including solitary rugose corals, trepostome bryozoans, etc., the dominant forms include small longicones (baltoceratids, *Cartersoceras popei*, and *Murrayoceras*), orthocerids (*Pojetoceras floweri* and *Proteoceras*), as well as the large actinocerid (*Actinoceras*). Also represented is the large endocerid (*Vaginoceras*), and a number of other unique forms. Many of these forms are holdovers from the Chazy of New York. As suggested by Ettensohn & Kuhnhenn (2002), the coiled nautiloid, *Plectoceras carletonense* is also represented but is usually fragmented and poorly preserved.

This high-diversity assemblage of nautiloids is important in that following deposition of the large K-bentonites and the widespread unconformity immediately above, nautiloids in North America suffer a significant and pronounced extinction (Frey, 1995). Following this event, only one form from the Tyrone survives into the Curdsville (*Vaginoceras* sp.). Thus beginning in the Curdsville and subsequent formations, the Trenton nautiloid faunas are characteristically species-poor (only four are known from the Curdsville) and nowhere are they as diverse as in previous assemblages. Although some taxa are locally abundant, similar patterns have been noted outside the region as well, including in the Decorah of Wisconsin-Minnesota, the Rockland of Ottawa, and the Watertown and Napanee of New York. As elsewhere some species become abundant on particular bedding planes in the Lexington, nonetheless, their diversity doesn't recover until new forms migrate into the region during the Cincinnatian. In terms of correlation outside the region, Frey (1995) indicated that the assemblage observed in the Tyrone was strongly characteristic of forms found in the Platteville Limestones of Minnesota and Wisconsin (11 of 14 species in common), the "Black River" of Lake Huron area of western Ontario (8 of 14 species in common), the Black River Group of the Black River Valley in New York (8 of 14 species in common), and the "Leray-Rockland" of the Ottawa area (12 of 14 species common). In the type New York region only a few taxa in the upper Black River (Watertown) are holdovers from the underlying Lowville, and none of these "survivors" carryover into the overlying Napanee. Frey (1995) thus suggests that the Kentucky Tyrone fauna is strongly a "Black Riveran" fauna, but that it could range upward to basal Rocklandian in age. Given this data and pattern of outage, it is possible that the nautiloid extinction noted by Frey (1995) and other previous workers may have occurred in two phases – the first phase may have occurred due to shallowing and emergence at the end of the Tyrone followed by a later extirpation during the ensuing transgression leading into the first Trenton highstand phase.

In addition to nautiloids, Yochelson (1966) reported a number of silicified gastropods from the Tyrone just about 15 centimeters below the Deicke (Pencil Cave) K-bentonite near the Logana type section in Jessamine County. In the acid digested fraction, in addition to significant amounts of quartz silt, Yochelson recognized the faunas to include the gastropods: *Helicotoma planulatoides*, *Trochonema*, *Loxoplocus*, *Lophospira*, as well as amphineuran plates, and opercula of *Euomphalacea*. *H. planulatoides* is a gastropod recognized in the Carters Limestone of Tennessee where it is associated with a much more diverse community (Wilson, 1949). In addition to the gastropods, another specific form has also been recognized from the mud-cracked facies of the Herrington Lake shale interval near the top of the Tyrone. At Camp Nelson a specimen of the Trilobite *Bathyurus extans* was collected from nearly identical facies to its occurrence in the top of the Lowville Formation of New York. According to Sloan (1987, 1991), *Bathyurus extans* and a number of other trilobites became extinct during this same time frame that is just before the large-scale transgression that ushered in Trenton deposition.

Trenton Group: Lexington Limestone General Description, Contacts & Distribution

The Lexington Limestone of the Jessamine Dome region was first recognized as a distinct unit above the level of the High Bridge Group by Campbell (1898). In the southern Jessamine Dome region, the thin-bedded gray limestone-bearing unit was originally defined as up to 48 meters of strata up to the level of the Flanagan chert **(see figure 3)**. Since originally defined, many revisions have taken place, including those of Black and colleagues (1965). The latter work established the precedent that the Lexington Limestone would extend upward to the base of the Clays Ferry Formation (to the top of the Cynthiana of early workers). Thus, this would include all strata up to the top of the Nicholas member of the Cynthiana (now Nicholas sub-member of the Tanglewood Member). In northern Kentucky and southwestern Ohio correlation of this unit is problematic as the coarse-grained Nicholas sub-member correlates northward into the interbedded shales and coarse-grained limestones of the Point Pleasant Limestone (McLaughlin et al., 2004). The latter unit is considered its own formation by some workers (Schumacher & Carlton, 1991) or as suggested by Weir and colleagues (1965) is a tongue of the Clays Ferry Formation. In northern Kentucky, McLaughlin and colleagues (2004) considered the Point Pleasant and underlying Bromley Shale as members of the Clays Ferry following the precedence of earlier workers and recognizing the original distinction of what was called the Cynthiana. In the Lexington region, Brett and colleagues (2004) included the Point

Pleasant as a formation within the Lexington Limestone – again reflective of the regional stratigraphic terminology.

This study follows the model proposed for the Lexington region originally by Cressman (1973) and used by Brett and colleagues (2004). Thus the Lexington Group would encompass all limestone-dominated strata up to the base of the Kope Formation of Ohio and would encompass the approximate interval represented by the Trenton Group of New York. Under the Black and colleagues model, which was modified slightly by Ettensohn (1992), the Lexington Group includes the Curdsville, Logana, Grier, Perryville, Brannon, Sulphur Well, Tanglewood, Devils Hollow, Millersburg, and Nicholas Limestone Members (**see figure 3**) as distinct lithologic units deposited on an irregular and complexly developed Jessamine Dome, referred to by Ettensohn as the Tanglewood "build-up." Cressman (1973) recognized all of these units, and a number of internal sub-units in each. However, Cressman favored the use of the term Point Pleasant in the Jessamine Dome region for the uppermost Lexington (Nicholas Limestone) – a precedent followed herein. In Ohio, the Point Pleasant is still considered as a distinct unit above the level of the Lexington Limestone (Shrake et al., 1990; Schumacher & Carlton, 1991) although the use of the term Point Pleasant should be brought into alignment with the former concept to avoid future complications.

As defined the Lexington Limestone is dominated by a fairly broad spectrum of limestone facies arranged in seemingly complex patterns. Black and colleagues (1965) describe the Lexington as a heterogeneous succession dominated by relatively coarse-grained bioclastic limestones, finer-grained fossiliferous limestones, and minor shales. Cressman reports the total thickness of the Lexington (from the Curdsville to the top of the Point Pleasant equivalent) to range up to nearly 100 meters of strata. Based on cores and outcrop studies, McLaughlin and colleagues report a maximum thickness approaching 110 meters for the entire interval. The Lexington is thickest in a linear belt oriented roughly east-west in the region just north and east of the Lexington area and the Kentucky River. The Lexington Limestones are shown to decrease in thickness toward the north, west, and southwest of this depocentral belt. This thinning is reported by Cressman (1973) to be the result of inter-tonguing and facies change of the upper Lexington with portions of the lower Clays Ferry Formation – especially in the vicinity of the margins of the Sebree Trough and to the southeast.

Distribution

The Lexington Limestone has been recognized as a fairly widespread unit in the Jessamine Dome and is well-exposed in the "inner Blue Grass" region of the Jessamine Dome. Black and colleague (1965) proposed reference sections along Interstate Highway I-64 just east of the Kentucky River, and additional sections are found along numerous highway cuts on I-75 in the Lexington region. The Lexington has been correlated southward through the Cumberland Saddle to equivalent strata in the Nashville Dome, (Holland & Patzkowsky, 1996), but even in early studies – the rocks of the Lexington area were equated with those of the Nashville Dome region (**see figure3**). In the subsurface to the east, the Lexington Limestone has been correlated by Pope (1995), Pope and Read (1997 a, b, 1998), and Pope and colleagues (1997) with the lower to middle Martinsburg Formation of Virginia. As suggested by wireline logs, the Lexington Limestone persists as a limestone-dominated unit at least as far as Pike County, Kentucky on the border with western Virginia and southwestern West Virginia. Immediately to the east in western Virginia and West Virginia, the Lexington Limestone transitions into the Nealmont and Dolly Ridge Limestones as recognized by Perry (1972) before transitioning into the Martinsburg Formation of north-central Virginia and eastern West Virginia.

To the north of the Jessamine Dome, the Lexington Limestone extends into the subsurface along the eastern axis of the Cincinnati Arch into the area referred to as the Point Pleasant Basin. Just north of Cincinnati, in the vicinity of Butler County, on the western axis of the Cincinnati Arch, the Lexington grades into a shale-dominated unit that has been referred to as the "Utica Shale" by Bergström and Mitchell (1990). The shale-dominated interval occurs in a northeast-southwest trending linear trough referred to as the Sebree Trough (Kolata et al., 2001). The Sebree Trough intersects the Cincinnati Arch in the central portion of western Ohio and apparently extends northeastward along the southeastern flank of the Findlay Arch to about the position of the Ohio/Pennsylvania border. In this region, near Lake Erie, Wagner (1966) recognized the same feature and evidence for a lithologic connection with the Sebree Trough. It is not clear if this feature continued northeasterly into the Lake Ontario Basin, but it does align approximately with the feature recognized as the Kingston Trough in the northeastern Lake Ontario Basin.

Across the Sebree Trough, in northwestern Ohio, Wickstrom and colleagues (1992) proposed that the entire interval of rocks above the basal Lexington (the Curdsville Member) up through the base of the Cincinnati Group (Kope Formation) be included in the Point Pleasant Formation. This interpretation was favored by the former authors as unlike in the Jessamine Dome, the Lexington Limestone equivalents grade into relatively homogeneous section of shales and limestones below the level of the Kope Formation. It was the interpretation of these authors that the Logana, Grier and subsequent units of the Lexington correlate into the succession they referred to as the Point Pleasant. Clearly, the use of Pt. Pleasant in this region is confusing and far removed from the concept of the unit as originally defined. Despite lithologic similarities of the interval to the Pt. Pleasant, with a more detailed analysis it is possible to differentiate at least some units, similar to those recognized in the southern Ohio-northern Kentucky region. Nonetheless, they are clearly separated from those of the latter region by the shales of the Sebree Trough and should not be considered Pt. Pleasant. Nonetheless, these observations suggest that the Lexington Limestone was deposited on a fairly broad central platform (aka the Lexington Platform) bordered on the north and west by the Sebree Trough, and on the east by the margin of the advancing peripheral style foreland basin (Brett et al., 2004). To the southwest the Lexington Platform extends into similar facies deposited on the Nashville Dome before transitioning into deeper water to the south.

Lexington Limestone: Curdsville Member

The Curdsville Limestone was first named as a bed of the Lexington Limestone by Miller (1905) for the extremely fossiliferous and "crystalline limestones" found at the base of the unit. Although the type locality is located in Mercer County, Kentucky, the Curdsville is a widespread, distinct sheet carbonate that is exposed in many locations across the Jessamine Dome and in the subsurface throughout much of the region (**figure 13**). Across this entire region, the Curdsville has a fairly uniform thickness. Black and colleagues (1965) report an average thickness for the Curdsville at just over nine meters in the southern Jessamine Dome and Cressman (1973) reports a maximum thickness for the unit to be about 12 meters in the same region. To the east of Lexington, in the Clark County Core (KGS CK-15) it appears that there might be as much as 18 meters of Curdsville developed just east of the Kentucky River faults. Northward of Shakertown, the thickness of coarse-grained facies thins beginning in the area of Lawrenceburg, Kentucky. From this point northward through Frankfort, Kentucky, well-developed Logana facies are clearly developed above the Curdsville so that in outcrop and cores

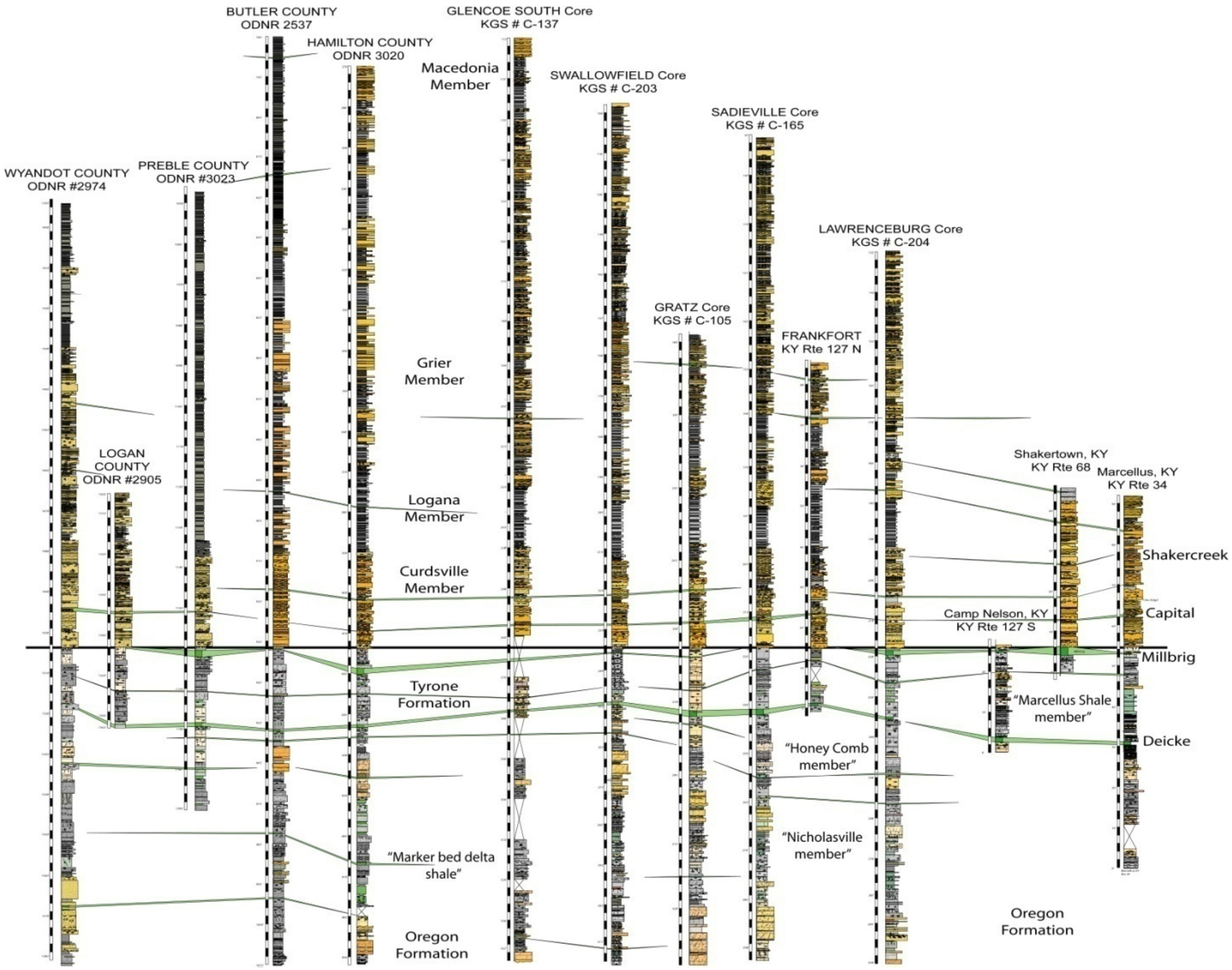


Figure 13: Correlated stratigraphic framework for the upper High Bridge Group to lower Lexington Limestone from the southern Jessamine Dome (Marcellus, Kentucky) northward to northern Ohio (Wyandot County). Correlations are based on numerous K-bentonite horizons (including the Deicke, Millbrig, Capitol, etc.) and the position of the Tyrone-Curdsville contact which is used as the datum.

the thickness of the Curdsville is only between six and seven meters for a distance of up to 350 kilometers.

Historically, Black and colleagues (1965) describe the Curdsville as a well-sorted bioclastic calcarenite that typically exhibits cross-bedding and chert-rich facies especially in the lower half of the unit. The limestones are also somewhat phosphatic and contain minor, thin, fine-grained micritic limestone beds and minor silty-shale partings that help to define small-scale cycles (**figure 14**). As mentioned, the Curdsville is developed between the High Bridge Group

Figure 14: Outcrop exposures and core photographs for the Curdsville Limestone. Outcrop photographs are taken from the KY-HWY 68 road cut at Shaker Creek, Kentucky and the reference section for the Shaker Creek K-bentonite. Core photographs are of the Ohio DCNR Core 3020 in Hamilton County, Ohio. The pattern and development of the Curdsville between these localities (some 100 km apart) shows remarkable similarities and relatively uniform thickness between these areas.

limestones below and the Logana Limestone above (**see figure 13**). At most localities, in the southern and southeastern Jessamine Dome south of Lexington and the Kentucky River, the Curdsville is overlain not by typical Logana lithologies (see below), but by lithologies similar to those of the overlying Grier (Cressman, 1973). This pattern is exhibited both in the Shakertown outcrop locality on KY Rte. 68, and at Marcellus on KY Rte. 34. The inability to separate the top of the Curdsville and the base of the Grier in this region – may explain some of the extra thickening of the Curdsville in this region. However, Ettensohn and Kuhnhenn (2002) suggest that these shallower facies are lateral up-dip, shallow-water equivalents of the deeper water Logana facies found to the north. If accurate, these shallow-water facies likely developed in response to a rather sudden change in the architecture of the Jessamine Dome during the onset of Vermontian tectonism and immediately after deposition of the Curdsville. Alternatively, the base of the Grier in this region could represent an unconformity that truncates the Logana down to the level of the Curdsville Limestone (see discussion below). Thus the increased thicknesses of coarse-grained Curdsville facies in the southern and eastern Kentucky River Fault zone is enigmatic and is likely tied to at least some syn-sedimentary movement on these faults.

The Curdsville was sub-divided by MacQuown (1967) into three sub-units, which Cressman (1973) indicates were informal and arbitrary in their assessment. Nonetheless, MacQuown recognized a lower unit which was 3 meters thick and composed of yellowish to light gray fine-to-coarse grained bioclastic calcirudites, cross-bedded and ripple-marked calcarenites, and laminated calcisiltites. Although dominantly calcium carbonate, the Curdsville shows a residue of between five and 10 percent fine quartz sand with additional components of disseminated chert and silicified skeletal grains (Cressman, 1973). Most skeletal grains, however, are highly abraded and sorted fragments of brachiopods, crinoids, bryozoans, and bivalves. The middle Curdsville unit was defined as another interval of up to three meters of bioclastic calcarenite and calcirudites (as below), but these become interbedded with argillaceous calcisiltites and dark carbonaceous shales. In contrast to the assessment of Cressman (1973), this change occurs at about the base of the Capitol metabentonite (of Conkin & Desari, 1986) – and can be recognized in many outcrops and cores readily. Minor chalky-white cherts are also developed in this layer and the rippled-calcarenites of this interval are often developed into hardgrounds with diverse encrusting communities attached to them. Finally, MacQuown (1967) recognized the upper unit (ranging from zero to upwards of six meters) as a medium gray irregularly bedded bioclastic calcarenite that can often develop into closely spaced brachiopod coquinas below the Logana. Cressman (1973) indicates that chert is rare in this part of the Curdsville and also indicated that this facies resembles that of the Grier member of the Lexington.

In addition to the prominent K-bentonites located in the underlying Tyrone Formation, the Curdsville also contains a number of K-bentonites recognized and correlated in local sections (MacQuown, 1967; Conkin & Desari, 1986; Kolata et al., 1996; Conkin & Conkin, 2000, etc.) In weathered outcrop sections most of these K-bentonites are not consistently recognizable and present difficulties for their use in correlation. However the use of portable scintilometers enables recognition of potential K-bentonite horizons, and the plethora of new road-cuts and rock cores has made it possible to recognize more of these horizons than previously assumed. As shown in figure 13 above, at least two prominent K-bentonites appear to be developed in the middle sub-unit. These include an "unnamed" K-bentonite and the Capitol in addition to the Shaker Creek of Conkin & Conkin (2000) – which is not recognized as frequently- even in core.

These two K-bentonites were correlated by Kolata and colleagues (1996) with the Elkport K-bentonite of the upper Mississippi Valley and the superjacent Dickeyville, respectively.

Despite the impact of major extinctions in nautiloid faunas at the end of the Tyrone (see discussion above), the Curdsville records a resurgence in the number of marine taxa – especially with respect to pelmatozoan, bryozoan, and brachiopod faunas. Numerous echinoderm forms have been recognized – especially from the middle unit of the Curdsville from the Mercer County region. The presence of these echinoderms, and their similarities with those of the Kirkfield Limestone of Ontario – helped suggest the Curdsville was Kirkfieldian in age. However many of the specific taxa are different. Several of the forms in the Curdsville are also similar to forms found in older rocks in the southern Appalachians and in the Upper Mississippi Valley, but again some are different. Thus the Curdsville fauna appears to be somewhat transitional between earlier and later forms. Included in the list of echinoderms of the Curdsville are: *Belemnocystites wetherbyi* –of which this specific clade is only known from the Bromide of Oklahoma and the Benbolt of Virginia (Parsley, 1972), *Carabocrinus vancortlandti* which is also found in the Decorah of Iowa (Brower, 1996), *Camarocystites tribrachius* (a paracrinoid form that is similar to *C. punctatus* that is found in the Ottawa region (Hull and Cobourg) of Ontario (Parsley, 1978), *Cremacrinus articulosus* (Brower, 1988), *Cupulocrinus humilus* (also found in the Snake Hill Formation of New York, the lowermost Hull Formation of Ontario, and the Decorah of Iowa), *Cupulocrinus jewetti* (known from the Platteville and Decorah of Iowa and the Hull and Verulam Limestones of Kirkfield and Ottawa), *Amygdalocystites florealis* with as many as two additional species possible (Bassler, 1915). At least twelve more genera from a number of classes are known. Branstrator (1979) also reports the occurrence of *Stenaster* cf. *S. obtusus* an asteroid species found in the Upper Ordovician through Lower Silurian. Especially productive regions are in the Garrard County region southwest of the Lexington. In this region, the echinoderm assemblages yield: edrioasteroids, paracrinoids, crinoids, and parablastoids.

In terms of bryozoans, early work by McFarlan and White (1948) suggested that the Curdsville contained a number of taxa as did the underlying High Bridge Group. Karklins (1983) indicated that the generic diversity of the lacy branching cryptostome (ptilodictyoid) bryozoans of Kentucky are similar to those of the Champlainian of New York and of the Decorah Shale of Minnesota. Although six genera were noted in these regions, four of these genera (*Stictopora, Trigonodictya, Escharopora*, and *Graptodictya*) are common between New York, Kentucky, and Minnesota. However, at the species level most forms in Kentucky appear to be endemic to Kentucky – that is these forms do not appear elsewhere. Forms recognized by Karklins (1983) in the Curdsville are generally long-ranging forms in Kentucky, but include *Orectodictya pansa* (only known from the Curdsville), *Trigonodictya cirrita, Stictopora neglecta*, *Escharpora eparmata*, and *Graptodictya sp*. – most all of which prefer shallow-water facies. Ross and Ross (2001) added the following forms: *Tarphophragma multitabulata*, and *Mesotrypa angularis*. These authors indicated that few Bryozoan species (mainly cryptostomes) were able to become established during the rapid sea-level rise of the Curdsville (and other transgressive events) – but suggested that highstands and later regressive phases have increased abundance and diversity of bryozoans.

In addition to the echinoderms and the bryozoans, other taxa recognized include one of the most dominant groups in the upper Curdsville – the brachiopods. McFarland and White (1948) and Cressman (1973) report up to fifteen different brachiopods including *Sowerbyella curdsvillensis, Dalmanella bassleri, Dalmanella fertilis, Hesperorthis tricernaria, Dinorthis pectinella*, *Rhynchotrema subtrigonale*, and others. Most brachiopods are usually highly abraded and fragmented in the lower Curdsville. However, upward, their preservation becomes somewhat better as many beds exhibit re-worked and winnowed brachiopod-dominated coquinal stringers. These are often filled with only one to two different forms at a time – usually however *Dalmanella* and *Sowerbyella* represent the most dominant forms found on bedding planes. Other taxa are somewhat less prevalent.

McFarlan and White (1948) also list the rugose coral *Streptelasma profundum* (also known from the underlying Tyrone of Kentucky, the Carters of Tennessee, the Platteville of Iowa, and the Lowville and Watertown Limestone of New York State). Other forms include *Brachiospongia digitata*, a number of gastropods including *Liospira sp., Lophospira sp., and Tropidodiscus sp.*, several lamellibranch bivalves (*Vanuxemia hayniana* and *V. umbonata, Cyrtodonta obesa and C. rotulata)* and a number of other molluscan forms (Miller, 1914). Cressman (1973) also listed the trilobites *Calyptaulax* cf. *C. strasburgensis*, and *Raymondites* cf. *R. ingalli* as occurring in the Curdsville Limestone. More recent field investigations have also yielded fragments of *Ceraurus* sp. *C. strasburgensis* is listed as occurring only in the Curdsville of Kentucky, from the Edinburg Limestone of Virginia (Brenner, 2004), and in rocks of the Northwest Territories of Canada of Black River age. Frey (1995) recognized the orthocerid *Anaspyroceras cylindricum*, a new species of *Vaginoceras* (Endocerid), and two actinocerids (*Actinoceras curdsvillense* and *Deiroceras curdsvillense*). Frey indicates that these are holdover taxa from the Black River and also indicates that none of these are found above the Curdsville.

Thus overall despite a major loss of large invertebrate taxa immediately preceding deposition of the Lexington Limestone, the faunas of the Curdsville record a few holdovers from the underlying High Bridge Group while numerous diminutive taxa appear in the region during the first major transgression of the Upper Ordovician. At this point, the Curdsville appears to retain a number of Black River taxa as well as carries the first representatives of the longer ranging Trenton taxa. Many forms appear to have been contributed to the Jessamine Dome from adjacent areas in the Upper Mississippi Valley, and the Nashville Dome – others still may have originated further north and west of the Transcontinental Arch – suggesting the arch was breached about this time.

Lexington Limestone: Logana Member

The Logana Member of the Lexington Limestone was first recognized as a distinct unit by Miller (1905) who recognized about three meters of strata immediately above the Curdsville Limestone in southern Jessamine County. In this region, the unit was described as an interval of extremely argillaceous limestones and very fossiliferous limestones immediately above the Curdsville. After initially recognizing the unit, Miller abandoned the term in favor of using the term Hermitage which had been correlated from Tennessee (Miller, 1913). In the southern Jessamine Dome, Miller applied the term Hermitage to between ten and twelve meters of thin-bedded fine-grained argillaceous and siliceous limestones and shales. An important component of this unit was the recognition of densely packed coquina beds containing *Dalmanella (Onniella) bassleri* and other orthid brachiopods. The top of the Hermitage was placed at the uppermost bed containing these brachiopods. Subsequent work reinstalled the term Logana in central Kentucky for this same approximate interval (Huffman, 1945). Cressman (1964) and Black and colleagues (1965), based on the challenges of mapping the Logana using faunal horizons rather than lithology, chose to redefine the Logana as a member of the Lexington Limestone. As a result of this work, the Logana was truncated below the top of the uppermost *D. bassleri* beds (which occurs in the overlying Grier) and the unit was redefined from the top of the Curdsville upward to the top of the uppermost interbedded micro-grained limestone (calcisiltite) and shale. With the contacts thus established the Logana was defined to contain up to about 9.5 meters of strata in the reference section (Black et al., 1965, Cressman, 1973; **see figure 13 above**). Along the Cincinnati Arch, the Logana approaches thicknesses up to 15 meters (near Falmouth, Kentucky; as reported by Cressman, 1973), but is most commonly 10 meters thick along this transect.

In the reference section for the Logana, the unit consists of fossiliferous limestones interbedded with between 40 to 50% shale (Black et al., 1965). In both outcrop and core, the unit is best defined as a calcisiltite and fossiliferous shale unit with occasional interbeds of tabular to wavy-bedded to lenticular coquinal beds (**figure 15**). The calcisiltites average between fifteen to twenty centimeters in thickness and are often gray to greenish gray in color on fresh surfaces. In weathered outcrops these beds most commonly weather to a yellowish-gray. Few beds of this lithology contain articulated fossils – and commonly exhibit uniform to slightly cross-bedded (HCS) textures suggesting rapid storm-influenced deposition. Nonetheless a few calcisiltite beds – especially at the base of the Logana are known to be fossiliferous and can be found to contain numerous articulated bivalves and *Cryptolithus* trilobite fragments. Some of these beds also contain disarticulated brachiopod (*Dalmanella*) stringers especially near the top of these beds (Cressman, 1973).

The intervening dark shales are often exceptionally carbonaceous – especially in fresh core, where they often impart a petroliferous odor. The shales are most often brownish-gray in color, and often weather to grey-white platy to nodular fabrics. Cressman (1973) reports that the shales of the Logana are dominantly composed of illite and chlorite-rich clays, as well as some minor quartz and affiliated accessory minerals. They are also rich in organics and often demonstrate mineralized surfaces (iron and phosphatic). In addition, the shale-dominated beds

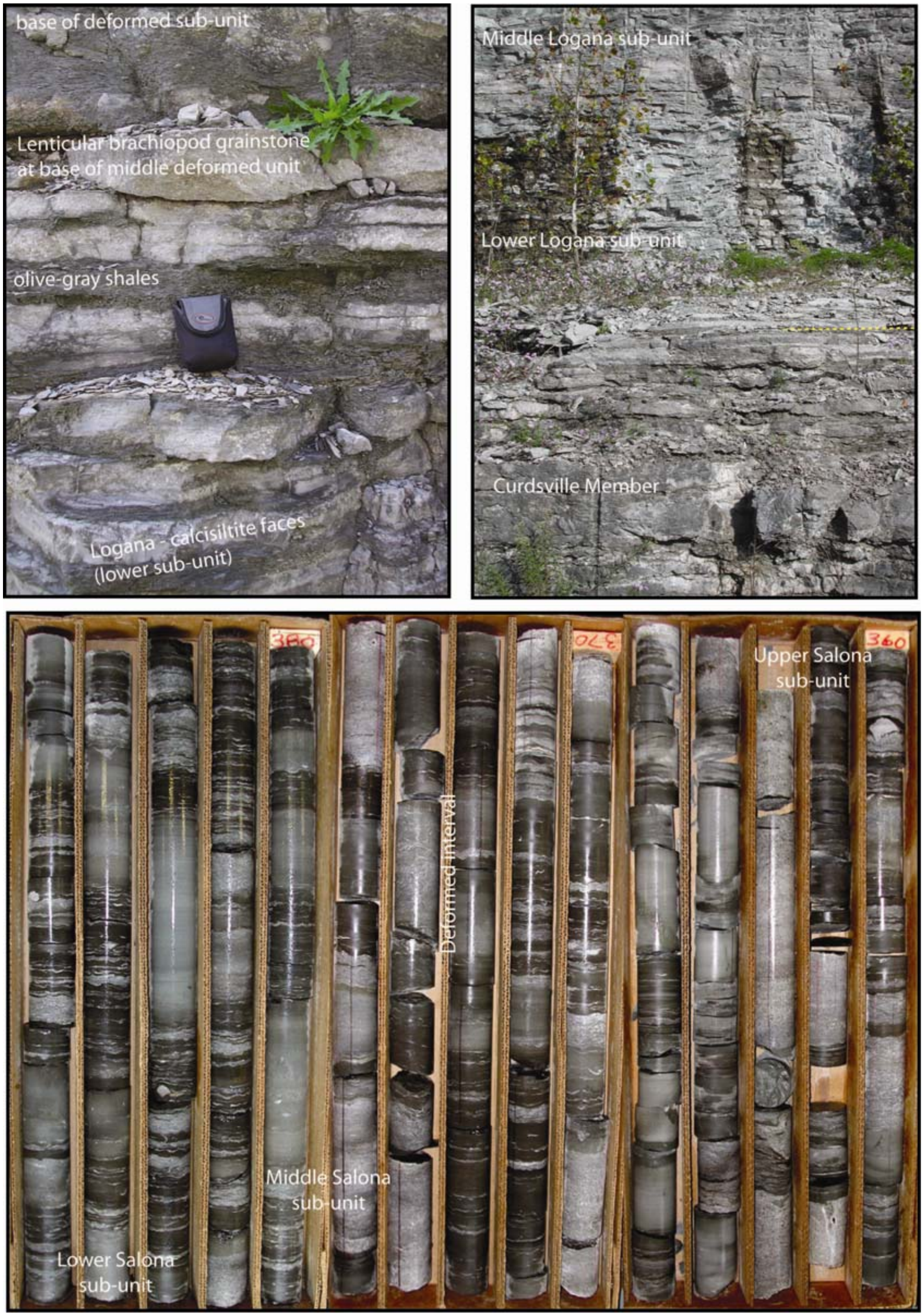


Figure 15: Outcrop and core photographs for the Logana Limestone Member of the Lexington Limestone. Outcrop photographs are from KY-HWY-127 north of Frankfort, Kentucky, while core images are taken from the Ohio DCNR Core 3020 in Hamilton County, Ohio. Photographs of the core represent the continuation of the images shown in figure 14 above – contact between the Curdsville and the Logana below is shown in figure 13 above. The key marker in the middle sub-unit of the Logana is shown.

contain numerous fossil components including *Dalmanella* brachiopods, bryozoans, and other fossils. In most cases, these forms appear to be in life position. Where the shales appear to have been disturbed, it is possible to observe coquinal pavements oriented with the concavity of shells oriented upward suggesting intermittent storm winnowing and settling in quieter waters. Some of these shale-hosted stringers can become much more closely stacked and or are welded to the bases or tops of calcisiltite horizons. This becomes especially predominant in the middle portion of the Logana. Cressman (1973) reports that the majority of the silt-sized grains in the calcisiltites were derived from crinoid and brachiopod material that was heavily fragmented and abraded and likely transported from shallow-water, winnowing regions some distance away – perhaps toward the south and east of the Jessamine Dome.

Like the underlying Curdsville, the Logana has been informally sub-divided into a three-fold unit that is generally recognized over most of the outcrop and cores in this study. The lower and upper sub-units are generally dominated by interbedded shales and barren calcisiltites, while the center sub-unit is dominated by coquinal packstones and grainstones through an interval of up to two meters. The middle sub-unit coquinas are interbedded with calcisiltites and shales. However, near the top of the middle sub-unit strata show evidence for soft-sediment deformation that can be recognized in core and outcrops. Although not as well-developed as subsequent deformed horizons – the occurrence of these deformed structures helps suggest ancestral faults in the Jessamine Dome region certainly became active during this time.

Cressman (1973) suggested that this middle Logana sub-unit could be correlated into sections to the south where typical Logana facies have not been recognized beneath the Grier. As mentioned, in these regions (i.e. Marcellus, Kentucky), the Grier Limestone appears to sit on the top of the Curdsville. In contrast to fairly uniform conditions across the region during deposition of the Curdsville Limestone, if Cressman's correlation is correct, then the recognition of the Logana deformed interval and associated grainstones above the typical Curdsville Member in the southern Jessamine Dome suggest that major topographic change occurred in the region at this time. It is likely that this "far-field" tectonic activity not only produced uplift of local portions of the Jessamine Dome, but may have also helped produce down-dropped areas as well. Specifically, it appears that the Sebree Trough first shows lithologic evidence for activation at this time. As shown in **figure 13** (Preble County, Ohio DCNR Core #3023), immediately above the Curdsville is an extremely condensed interval overlain by a succession of interbedded shale and calcisiltites – much like the Logana – except that these beds lack coquinal interbeds and are barren of major faunal elements. Collectively recognition of seismic events and the first occurrence of major lateral lithologic change suggest that following a period of quiescence at the end of the Blountian tectophase, this interval demarcates one of the first pulses of the Vermontian tectophase of the Taconic Orogeny.

Other marker horizons within the Logana are relatively minor and discontinuous. Unlike the underlying Curdsville, K-bentonites within the Logana are relatively discontinuous and appear to be absent in many cores as they do not stand out when wetted. At least two to three K-bentonites are present, but these are not consistently recognized and are therefore not described herein. Aside from the middle deformed marker horizon, the only additional stratigraphic marker within the Logana is that of the positive Guttenberg Isotopic Carbon Excursion (GICE). In the area of Frankfort, Kentucky the GICE has been recognized by Young and colleagues (2005) as beginning above the upper Curdsville and extending upward to the base of the Grier. The largest portion of the excursion occurs in the lower sub-unit before values begin their decline back to negative values at the base of the middle sub-unit. In the photographs shown in figure 15 above, the dark organic-rich shaly intervals below the middle sub-unit thus contain evidence for the GICE.

As mentioned, the Logana has a brachiopod-dominated fauna with many orthids including dalmanellids. Cressman (1973) reports at least three different dalmanellid taxa (*Dalmanella fertilis* and *Dalmanella sulcata*, as well as another un-named form). Other brachiopods include *Rafinesquina trentonensis, Rhynchotrema* sp., *Rostricellula minuta*, *Platystrophia colbiensis*, and *Zygospira sp*.. A molluscan fauna is also noted and includes: monoplacopherans (*Cyrtolites* aff. *C. retrorsus* now recognized as *Paracyrtolites subplanus* by Wahlman, 1992), gastropods (*Carinaropsis cymbula*, *Liospira* aff. *L. decipiens*, and *Lophospira* sp.), a number of bivalves (at least eight different forms), and a minimal nautiloid fauna. Frey (1995) show nautiloids in the Logana to be noticeable but of low diversity. Three recognizable species have been identified and include small longicones forms *Isorthoceras albersi* (most abundant), *Gorbyoceras* cf. *G. tetreauense*), and the orthocone *Allumettoceras* cf. *A. tenerum*. The majority of these forms are characteristic of the remainder of the Trenton and Lexington.

Also recognized are at least two ostracod taxa (*Ceratopsis intermedia* and *Jonesella* aff. *J. obscura*), as well as a couple of trilobite species (including *Cryptolithus tesselatus*; and an unspecified encrinurid; Ross, 1979). Other important taxa are the bryozoan groups that become important in some intervals and are especially prevalent in the shale interbeds of the Logana. Brown (1965) identified 20 species and one sub-species of trepostome bryozoans from the Logana (and the overlying Jessamine or Grier Member). In the outcrops studied, Brown was unable to differentiate these units lithologically, and based on the bryozoan assemblages considered them to be relatively similar. The key forms recognized belong to the following genera: *Amplexopora, Cyphotrypa, Dekayella, Dekayia, Eridotrypa, Hemiphragma, Homotrypa,* *Homotrypella, Hallopora, Lactopora, Prasopora, and Stigmatella.* Karklins (1983) identified only one ptilodictyoid species in the Logana which included *Graptodictya* which is also known from higher units in the Lexington, but the species is different from the form known from the underlying High Bridge Group. Thus given the fauna recognized in this stratigraphic interval, compared to the underlying Curdsville, the Logana is more allied faunally with strata of the overlying units rather than those of underlying successions (including the High Bridge Group and the Curdsville Limestone).

Lexington Limestone: Grier Member

The Grier Limestone member of the Lexington Group was named for exposures of this unit along Grier Creek in Woodford County, Kentucky and for exposures along Shyrock Ferry Road. Mapped by Cressman (1964) and raised to member status by Black and colleagues (1965), the Grier was defined as a light yellowish-gray, medium to coarse-grained, fossiliferous limestone. The unit is thin-bedded, often nodular, and slightly phosphatic, and contains numerous thin shale beds. The Grier is the fourth stratigraphic name applied to this specific interval of strata in central Kentucky. Initially, this interval was considered an equivalent of the Hermitage of the Nashville Dome region and was used by Miller (1905) in Kentucky for this specific interval and was used by McFarlan and White (1948) to include the Curdsville, Logana, and Jessamine Members). Preferring a local name, Nickles (1905) used the name Wilmore in place of Hermitage and other workers variously recognized this unit. The name Jessamine was eventually applied to this interval by Foerste (**see figure** 2), and was used by McFarlan (1931, 1938, 1943), and by McFarlan and White (1948). Unfortunately the Jessamine was defined as a biostratigraphic unit based on the zone of abundant and typical *Prasopora simulatrix* and *P. falesi* along with a number of brachiopods first observed in the underlying Logana. Given the apparent difficulty in mapping the Jessamine with confidence, Cressman (1964) proposed the name Grier to supersede any previous stratigraphic nomenclature. Cressman considered the unit to range up to about 35 meters in total thickness.

As defined by Cressman (1964) and Black and colleagues (1965), although the base of the Grier was placed at the base of the highest bed of the Logana, when the Grier-type facies was correlated to the south into Jessamine and Fayette Counties the typical Logana facies is not present below. In its place, facies more typical of the Grier extended downward to the top of the Curdsville Limestone. Thus, the base of the Grier was considered to be time transgressive in the southern Jessamine Dome region and the contact with the Curdsville was said to be gradational and intertongued across an interval of up to three meters (Cressman, 1973). Detailed studies in the central to northern Jessamine Dome by Brett and colleagues (2004) and McLaughlin and colleagues (2004) have helped recognize the contact to be prominent with substantially shallower facies immediately overlying deeper-water rhythmite facies. This contact is synchronous across the northern and central Jessamine Dome, and is correlated into the Grier as exposed at Marcellus, Kentucky, where it occurs at the base of a brachiopod grainstone and intraclastic breccia bed approximately six meters above the top of the Curdsville Limestone. As suggested by Cressman (1973), this six meter-thick interval correlates with the Logana and is older than the type Grier.

The upper contact of the Grier was originally defined and established at the base of the widespread and fairly uniformly-developed calcarenites of the Tanglewood Limestone Member of the Lexington (Black et al., 1965; Cressman, 1973). Ettensohn and colleagues (2002e) referred to this latter interval as the lower tongue of the Tanglewood Member of the Lexington Limestone and recognized it to sit immediately below the Brannon Member of the Lexington

Limestone. This tongue is equivalent to the interval referred to as the Benson by early workers or a part of the Perryville Limestone (McFarlan & White, 1948). Black and colleagues (1965) established the top of the Grier at the top of a two-meter thick argillaceous limestone some distance below a stromatoporoid-bearing interval in the Tanglewood Member. These authors applied the term “Cane Run Bed” to this interval and included it in the top of their Grier Member. Ettensohn and colleagues (2002e) attribute upwards of nearly seven meters to the “Cane Run” interval in outcrops south of the Kentucky River along KY HWY Rte. 27 where they have expanded the concept of the Cane Run to include coarsely crystalline grainstone beds, and argillaceous cherty limestones. In this location, the “Cane Run” demonstrates two distinct deformed intervals. However, McLaughlin and Brett (2006) suggest these “Cane Run” seismites are actually in the Brannon Member.

Nonetheless, Brett and colleagues (2004), McLaughlin and colleagues (2004), and McLaughlin and Brett (2007) recognize the upper Grier of older workers and the overlying lower Tanglewood tongue, as equivalents of the Perryville Member of the Lexington Limestone (after Nickles, 1905). The Perryville is made up of three sub-units: Faulconer, Salvisa, and Cornishville Beds. The Cornishville and the underlying Salvisa represent the stromatoporoid-bearing Tanglewood tongue (of Ettensohn et al., 2002 d,e) and the underlying Faulconer Bed is the equivalent of the uppermost Grier. Given the lithologic distinction of the Perryville Limestone and its sub-units from the type-Grier and the historical precedent of this former unit, it is suggested that the Faulconer Bed equivalent be excluded from the Grier and considered as a sub-unit of the Perryville Member as originally defined. In the McLaughlin and Brett (2007) model the distinct facies in the upper portion of the Grier referred to as the Macedonia Bed, is also excluded from the Grier as it is clearly a distinct and recognizable facies much like the

Logana, and the Brannon. Given the prominence of the Macedonia Bed it is herein considered as a distinct unit separate from the underlying Grier. Therefore the top of the Grier is placed at the base of the Macedonia Beds where the coarse-grained brachiopod and bryozoan coquinas give way upward to interbedded fossiliferous shales, calcisiltites, and minor stringers of packstone to grainstone lithologies similar to the underlying Logana Member. Thus as re-defined, the Grier is substantially thinner than its original concept so that it reaches a maximum thickness of just about 13 meters in the southern Jessamine Dome (Danville area) and has an average thickness of about 9 meters in exposures and cores along the Cincinnati Arch to the region of Cincinnati (Brett et al., 2004).

Exclusion of the uppermost interval of the Grier (as per the definition of Black et al., 1965), is significant and important in that it helps to narrow the Grier to a rather more uniform set of lithologies. Moreover, this effectively brings the modified Grier into alignment with the descriptions of most workers. For instance, Wahlman (1992) described the Grier as consisting of thin, irregularly bedded to nodular-bedded, poorly sorted, fossiliferous limestones that were relatively rich in brachiopods, bryozoans, gastropods, ostracodes, crinoids, bivalves, and trilobites.

As originally defined the Grier was sub-divided into four units that were useful in correlation. Cressman (1973) indicated that in addition to the formalized Macedonia and Cane Run Beds another two units (lithofacies) were recognizable and considered informally. As two of these occurred above the Macedonia they are not discussed here. However, the lower sub-unit recognized by Black and colleagues (1965) was recognized on the basis of an interval of approximately one meter of bryozoan-rich limestones with very large crinoid columnals (likely *Cleiocrinus*) that was superseded by up to 3 meters of lenticular to nodular-bedded brachiopod packstone coquinas. Cressman indicated that these marker intervals occurred approximately three to four meters below the Macedonia Bed. Collectively these units were correlated over much of the southern Jessamine Dome and are approximately equivalent to the middle sub-unit of the Grier as defined here. Thus as now recognized by McLaughlin and colleagues (2004) and McLaughlin and Brett (2007), the Grier can be subdivided into three sub-units – a lower, middle, and upper – all below the level of the Macedonia Beds (**figure 16).**

The lower unit is predominantly composed of shaly nodular wacke to packstone overlain by an interval of calcarenite facies (in the area around Marcellus, Kentucky) and grades into skeletal grainstones and rudstones to the north as exhibited in the outcrops around Frankfort. The middle sub-unit of the Grier is typically developed as a shaly nodular facies that is typically dominated by a diverse fauna including the trepostome bryozoan *Prasopora* and may contain large fragments of crinoids – the lower sub-unit recognized by Cressman (1973). Northward of Frankfort, the middle sub-unit of the Grier maintains a fairly uniform appearance through the Cincinnati region where it begins to transition into a more rhythmically-bedded facies, but is still recognized below the upper Grier (**see figure 16**). The uppermost sub-unit of the Grier is composed of up to an average of three meters of skeletal calcarenite and rudstone facies immediately below the level of the Macedonia Beds in the central Jessamine Dome. Northward the unit thickens slightly to just less than five meters. Toward the south, the upper Grier can be traced into shallow-water (peritidal) facies in the region around Danville suggesting this region was significantly shallower than the rest of the Jessamine Dome region (McLaughlin & Brett, 2007).

Although few specific marker beds are recognized in the Grier, beyond that already discussed, the Grier is somewhat distinct in that it typically has an abundance of phosphate.

Figure 16: Core photographs of the re-defined Grier Member of the Lexington Limestone. Core photographs are from Swallowfield area north of Frankfort Kentucky, KGS Core # 203. As shown is the tripartite nature of the Grier from its basal contact with the Logana up through the contact of the Grier with the overlying Macedonia Beds. Although in weathered outcrop where the Grier appears to be a rather monotonous uniform lithology, it is appear to be very cyclical as shown here.

Cressman (1973) indicated that the Grier on average has several times more phosphate than other limestones in the region. In the Grier, the phosphate occurs as disseminated cryptocrystalline forms usually found as infillings in the pores of bryozoan zooecia, in crinoid plates, and in other fossils where it forms coatings on shells, but only rarely replaces minor portions of skeletal remains. It is surmised that much of this phosphate is the result of very early bacteria-mediated diagenesis just on or under the sea-floor.

The Grier shows a significantly higher diversity fauna than the underlying Logana. Cressman (1973) reports a substantial list of taxa for the Grier, exclusive of the Cane Run and Macedonia Beds. The unit is documented to contain corals (including *Tetradium* sp., and *Favistina* cf. *F. stellata*), nearly twenty or more bryozoa taxa including the characteristic *Prasopora* (*P. falesi* and *P. simulatrix*), and the first appearance of *Constellaria* cf. *C. teres*. Also recognized are nearly fifteen typical lower-middle Trenton brachiopod taxa, a wide range of molluscan faunas including gastropods, monoplacophorans, bivalves, and nautiloids. Cressman (1973) also denotes an increased diversity of arthropod taxa including at least four different species of trilobites (including *Flexicalymene*, *Gravicalymene*, *Isotelus gigas*, and *Proetidella*? sp.) and at least nine species of ostracods.

As suggested by Frey (1995), the faunas of the Grier (and the nautiloids in particular) are characteristic of the Trenton of New York as described by Titus and Cameron (1976). The nautiloids of the Kings Falls to Sugar River match these remarkably well. These are particularly abundant in the uppermost sub-unit of the Grier immediately below the Macedonia and again in the interval immediately overlying the Macedonia (as per Ettensohn et al., 2002). Clearly, compared to previous stratigraphic intervals, the Grier represents the first major incursion of typical, diverse, Trenton taxa – although in most cases preservation is relatively poor owing to significantly more winnowing and wave reworking – especially in the lower Grier. The occurrence of the large crinoid columnals is also intriguing from the middle sub-unit. Large crinoid columnals with large lumens are typically derived from the genus *Cleiocrinus* and are found in abundance in some beds of the Kirkfield Limestones and occasionally are found in the Kings Falls Limestone of New York (Titus, 1986). In this regard, and with recognition of the GICE in the subjacent Logana, at least a portion of the Grier is coeval with the Kings Falls and Kirkfield of the type region. Moreover, the abundance of bryozoans and brachiopods etc., in the middle to upper Grier suggest Shermanian equivalency for these units.

Lexington Limestone: Macedonia Beds

As mentioned the Macedonia Beds were established as a distinct interval of "Brannon-" or "Logana"-like facies in the middle of the Grier Formation (sensu Black et al., 1965). This interval was included as a sub-unit of the Grier, presumably due to its relatively thin nature (ranging from about two to five meters in total thickness). Previously, McFarlan and White (1948) recognized this specific interval in the bottom of their Benson Limestone. Therefore given its long-running distinction, it is included here as a distinct unit on the grounds that the Macedonia is a widespread interval that is easily recognized in central to northern portions of the Jessamine Dome and northward into northern Kentucky to southwestern Ohio where it transitions into shale-dominated strata. The Macedonia is an interval of tabular to lenticular-bedded argillaceous calcisiltites and interbedded shales that occasionally contain minor fine-grained calcarenite stringers – especially in the southern Jessamine Dome. Interestingly, in the southernmost Jessamine Dome, the Macedonia was reported to contain *Tetradium* and ostracods (Ettensohn et al., 2002 d,e). In addition, these authors identified hummocky cross-stratification and minor graded beds developed in its upper part where it grades upward into the Faulconer sub-Member of the overlying Perryville, which is also known for these corals. Northward, the

Macedonia is predominantly a shaly nodular facies that becomes dominated by rhythmite facies in the central Jessamine Dome (Frankfort region) and further north in the area of Cincinnati the Macedonia transitions into laminated shales with only minor calcisiltite interbeds.

The Macedonia can also be described as a tripartite unit (**figure 17**). Immediately overlying the coarse skeletal grainstones and rudstones of the underlying Grier, the first homogeneous calcisiltites interbedded with calcareous shales appear. This succession of lower Macedonia beds appears to grade upward into a somewhat shallower and coarser-grained skeletal wackestone to rudstone (middle Macedonia beds) which in turn are overlain by another succession of calcisiltites and shales (upper Macedonia beds) that often contain cherts and a K-bentonite near the top. This K-bentonite is intermittently persistent immediately below the contact with the Perryville and has been referred to as the Westboro K-bentonite (Brett et al., 2004) after correlation of this interval from Ohio where the Westboro was first recognized (Schumacher & Carlton, 1991). The top of the Macedonia is established at the base of the somewhat shallower and more nodular facies. This contact is roughly the upper contact of what was referred to as the Jessamine Limestone of Nosow & McFarlan (1960), and is referred to as the Faulconer Bed of the Perryville Member of the Lexington Limestone (McLaughlin et al., 2004; McLaughlin & Brett, 2007).

Faunally, the Macedonia has also been characterized by previous workers and a detailed list was published by Cressman (1973). The Macedonia is a bryozoan-brachiopod dominated unit much like the Brannon (see below). Cressman includes *Eridotrypa*, *Heterotrypa*, *Peronopora* cf. *P. granulifera* and *Prasopora falesi* (now considered a synonym of *P. simulatrix*) as the main bryozoan taxa. He also recognized the brachiopods *Hebertella frankfortensis*, *Heterorthina macfarlani* and *Rafinesquina trentonensis* as well as the trilobites *Flexicalymene*

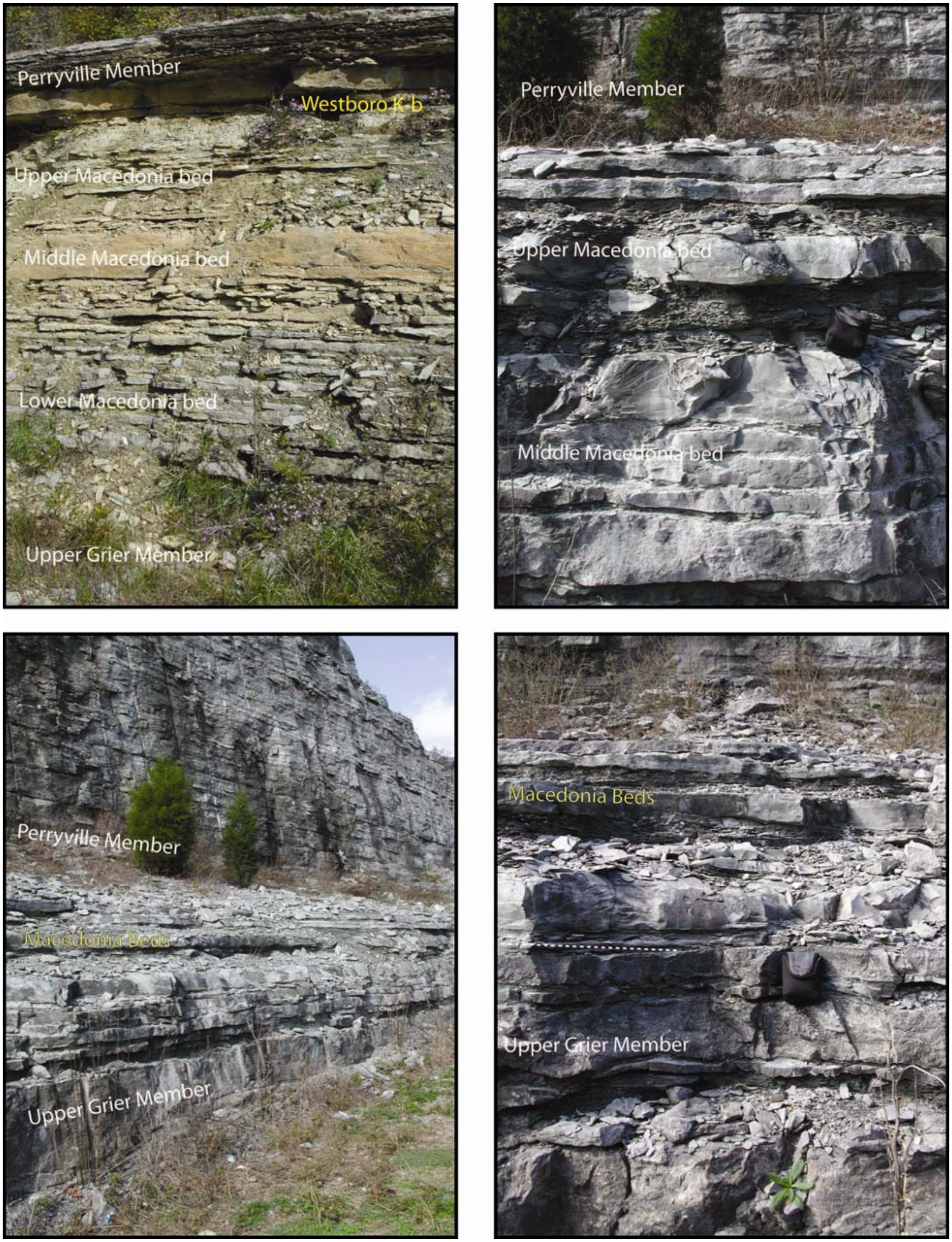


Figure 17: Outcrop photographs of the Macedonia interval as exposed in the vicinity of Frankfort (KY Rte 421 – upper left; and KY Rte 127 – bottom left and right side images). Clearly shown is the tripartite nature of the Macedonia and the typical bedding style and associated lithologies. The position of the Westboro K-bentonite is also established near the top.

sp. and *Proetidella* sp. which is a synonym of *Decoroproetus* (Ross, 1979). Ross indicated that there are two un-described species of *Decoroproetus* in this approximate interval (through the top of the Perryville). Aside from one ostracod taxa no other forms are listed.

Lexington Limestone: Perryville Member

The unit immediately overlying the Macedonia Beds has been considered to be part of the upper Grier Member of the Lexington Limestone as defined by Black and colleagues (1965). By these authors, the Grier is inclusive of the units up to the base of the Brannon Member and the contact with the overlying Cynthiana Formation of older workers (i.e. see Nosow & McFarlan, 1960). The relationship between the Grier and overlying units has been considered to be complexly inter-tonguing and laterally variable (Cressman, 1973; **figure 18**). Nonetheless the package above the Macedonia is faunally and lithologically distinct from the underlying Grier at least in the southern outcrop region. In this region it is referred to as the Perryville Member of the Lexington. It displays a number of distinct lithologies and shallower-water taxa and thus has been separated on that basis from the Grier by previous authors.

North of Perryville in the Frankfort, Kentucky area, McFarlan & White (1948) recognized the same interval as the Benson Member of the Lexington Limestone and divided it into an upper and lower unit. They considered the upper to be the equivalent to the Perryville Limestone to the south. The Benson was considerably less shaly than the underlying Jessamine and was in places fairly heavy-bedded and coarser grained. The unit was easy to recognize because of the incursion of *Stromatocerium pustulosum*, *Dinorthis ulrichi*, *Strophomena vicina*, and *Cyphotrypa frankfortensis*, which were especially abundant in the upper part of the unit. In

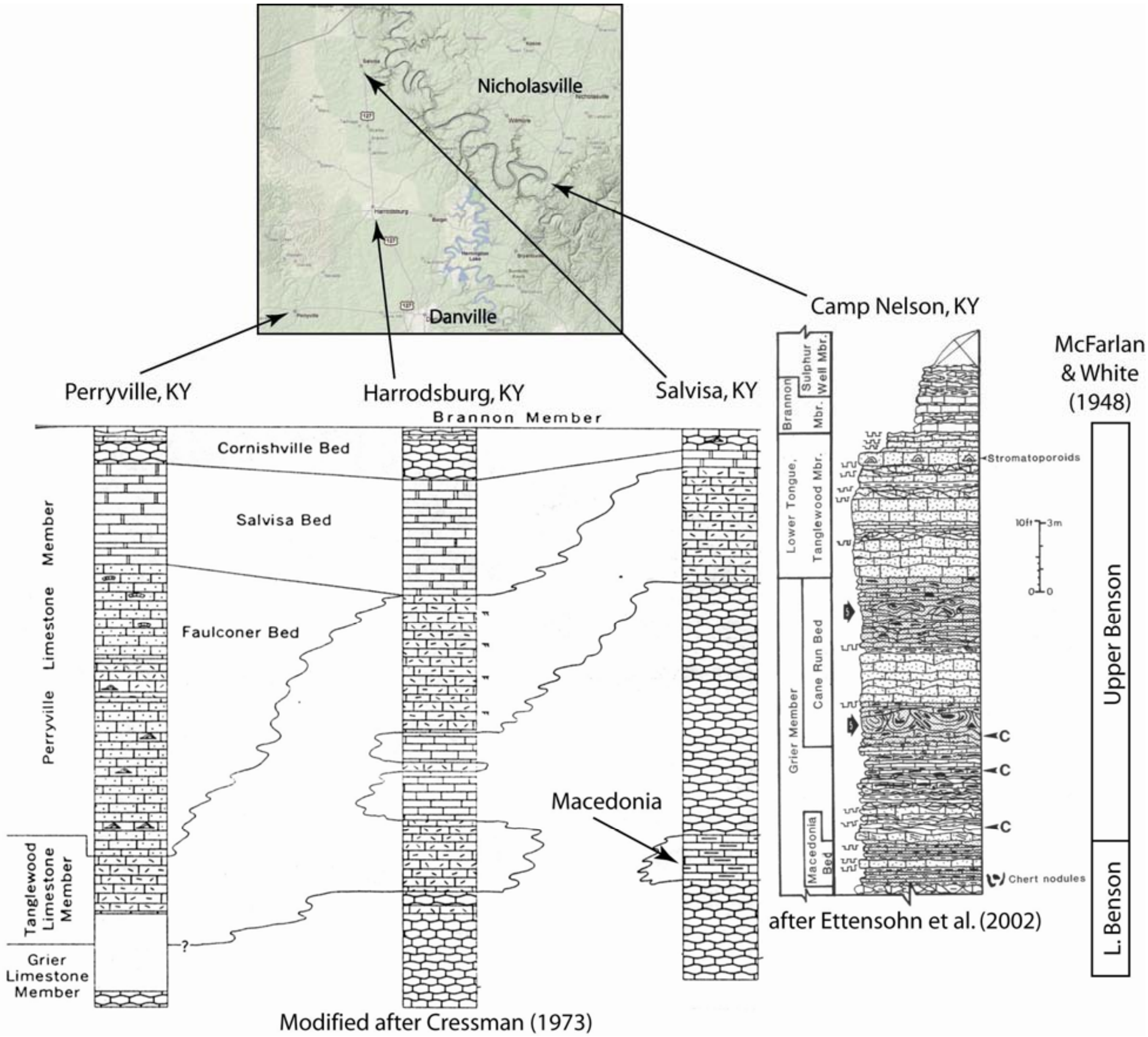


Figure 18: Review of stratigraphic nomenclature as applied by Cressman (1973) and Ettensohn et al. (2002) for the interval beneath the Brannon Member of the Lexington Limestone. The stratigraphic section of Cressman forms a roughly south to north transect, and the Ettensohn et al. column applies for regions to the south of Nicholasville near the Kentucky River. Also shown is the approximate interval referred to as the Benson by McFarlan & White (1948).

the type section of the Brannon (Jessamine County-Kentucky), Nosow and McFarlan (1960) identify an interval of at least 5 meters of strata below the Brannon as their upper Benson interval. The Benson continues downward some distance, but the base of the unit was not specifically defined due to lack of exposure. However, in this exposure interval a *Stromatocerium* biostrome unit is identified and can be used as a widespread marker horizon and correlated elsewhere.

As mentioned, in the region near Danville, Kentucky, the Benson transitions into the Perryville Limestone (Nosow & McFarlan, 1960; **see figure 18**). The Perryville was first recognized by Nickles (1905), and was subsequently subdivided into three units: the Faulconer (Foerste, 1912), the Salvisa (Miller, 1913), and the uppermost Cornishville (Foerste, 1912). The uppermost of these units, the Cornishville bed, contains an interval of abundant *Stromatocerium* (the biostrome marker of the Benson) and its associated fauna (**figure 19**). The underlying Salvisa and Faulconer sub-units, although also containing stromatoporoids and *Tetradium* corals, are equivalent to and correlated with lower portions of the Benson further north. As suggested by McFarlan and White (1948), the Perryville is a Benson facies [or vice versa], and at least in the north is similar to typical Lexington. Given this stratigraphic relationship, the Perryville has precedence over the term Benson (although it is a shallower-water equivalent), and hence has been used by McLaughlin and colleagues (2004), McLaughlin and Brett (2007), and herein for the interval immediately above the Macedonia and extending upward to the base of the Brannon.

Although the Perryville was not recognized as a widespread unit by Black and colleagues (1965), it can be correlated into more offshore facies to the north. Lithologically, these authors considered the Perryville to be a fine-grained, fossiliferous calcilutite facies, interbedded with coarser calcarenites and relatively barren calcilutites. These lithologies were located immediately above their lowest tongue of Tanglewood calcarenites (**see figure 18**). However, as originally defined, the Perryville contains a tripartite succession. Unfortunately north of the type region, the base of the Perryville equivalent (Benson of Nosow and McFarlan, 1960) is covered in most outcrops, and only the upper portion was studied in detail by these authors – which may have contributed to some confusion. Nonetheless, the three sub-units of the Perryville Limestone Member were described as follows for exposures in the Danville region (McFarlan &

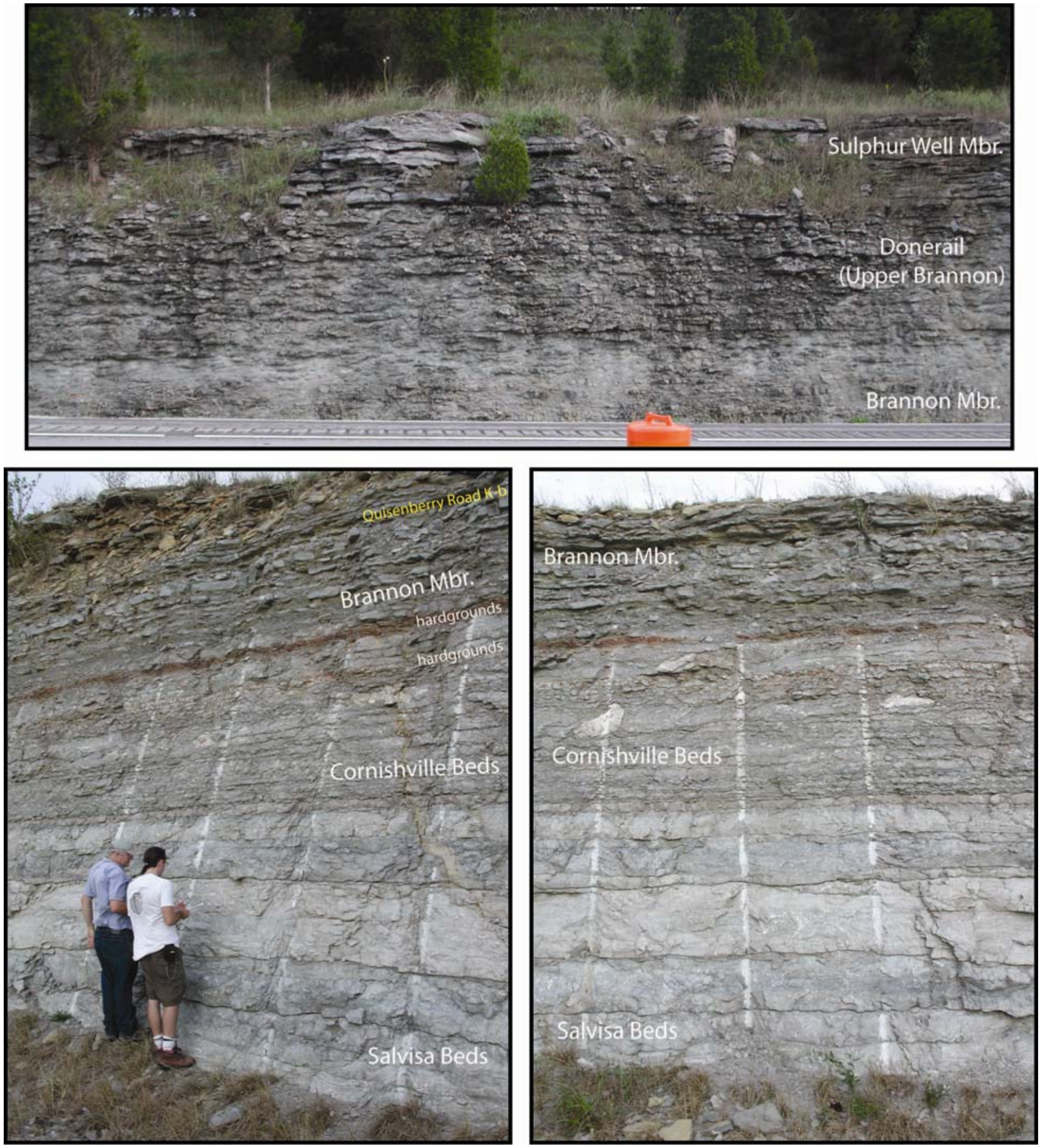


Figure 19: Outcrop photographs of the upper Perryville and Brannon Members of the Lexington Limestone in the Danville, Kentucky region. Clearly shown are the relatively thick-bedded fine- to coarse-grained facies of the Salvisa and the thinner nodular facies of the Cornishville Beds. Note the presence of overturned *Stromatocerium* in some levels of the Cornishville. Top of the Cornishville is marked by a pair of closely-spaced hardgrounds immediately below the Brannon Member. The Brannon is in turn overlain by the Donerail and the Sulphur Well Member (above). In this locality, the upper Brannon seismites are developed further west in this outcrop although they are not shown in the image above.

White, 1948; **figure 20**). The Faulconer Bed was considered to be a rather massive, light gray, coarsely crystalline, fossiliferous limestone that ranged from 0 to about 4.5 meters in thickness near Danville owing to facies change into what was called "upper Grier." Cressman (1973)

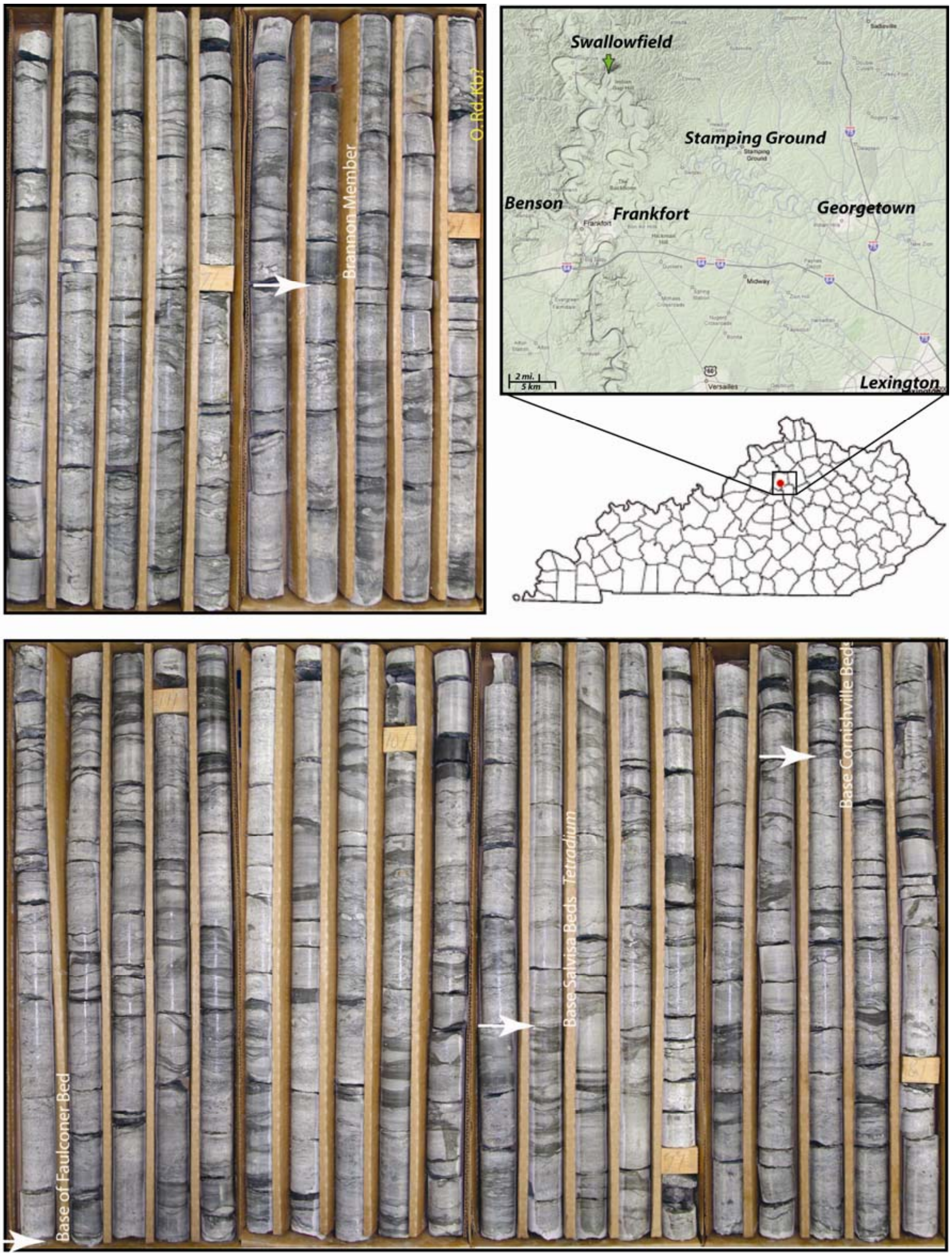


Figure 20: Core photographs for the Perryville Member of the Lexington Limestone. Core is KGS C-203 from Swallowfield, Kentucky in northern Franklin County. This interval was included as uppermost Grier by Black et al., 1965, however individual sub-units of the Perryville are recognizable to the north of their type region. Core shows very base of Perryville (Faulconer Bed) above the level of the upper Macedonia, through the middle (Salvisa Bed) and upper (Cornishville Bed) sub-units of the Perryville where it contacts the overlying Brannon Member of the Lexington.

attributes a total of about 12 meters for this bed in the Perryville area.

The Faulconer Bed is in turn overlain by the Salvisa (another 0 to 4.5 meter thick interval) which was considered to be a fine-grained calcilutite facies that contained some "birdseye" (fenestral) micrite especially in southernmost exposures. Cressman (1973) reports a total of just less than 5 meters for this interval at Perryville and Harrodsburg, Kentucky. The upper unit, the Cornishville Bed, was again characterized by a more typical "Lexington" lithology that was fossiliferous and variously interbedded with coarse- to fine-grained interbeds reaching a thickness of up to 2.5 meters. This latter unit was especially important as it contained the "upper Benson" stromatoporoid and associated faunas found further north. Thus, in total, in the Perryville type region, the thickness of the unit ranges up to a total of 11.5 meters (McFarlan & White, 1948). At Perryville, where the upper Grier is covered, Cressman (1973) reports a total of nearly 19 meters and only a portion of that are shown to be present north of Salvisa (only a few meters of uppermost Perryville are recognized as upper Salvisa and Cornishville).

Using the Cornishville and its faunas for correlation to the north, McFarlan and White (1948) however, suggested the Salvisa equivalent was present although it became a more compact limestone unit that was interbedded with thin shales. As in the south, the unit is characterized by an abundance of ostracods and associated brachiopod taxa. With regard to the underlying Faulconer, these authors recognized the unit as a relatively coarse-grained interval with moderately thick beds, but also weathering to a nodular appearance in some places. In the region to the north of Salvisa, the Faulconer is also interbedded with minor shales and contains thin nodules of chert. The unit is also characterized by large numbers of gastropods including bellerophonts, and other mollusks. These are often preserved by silicification by white - gray cherts.

Correlation of Cressman's (1973) stratigraphy to the east into the area of Camp Nelson southwest of Nicholasville is shown in **figure 18**. Ettensohn and colleagues (2002) recognized an interval at Camp Nelson above what they interpreted as the Macedonia Bed. This interval shows an increased abundance of chert nodules and cephalopods that are preferentially oriented – evidently by bottom currents and wave activity. Recognition of: 1) a Macedonia-like facies, 2) a succession of twin deformed intervals, and 3) a stromatoporoid-bearing unit suggested to Ettensohn and colleagues (2002) and Jewell and Ettensohn (2004) that this section was the equivalent of the upper Grier (after Black et al., 1965). In their model, the stromatoporoid unit was considered to be the Cornishville Bed below the Brannon, and the deformed intervals were considered to be the Cane Run Beds of the upper Grier. Based on their correlations the cherty, cephalopod-rich interval (of Ettensohn et al. 2002) is positioned in approximately the same position as the lower silicified, mollusk-bearing, Faulconer Bed to the southwest. In contrast, investigations of the Camp Nelson section near the Kentucky River by McLaughlin & Brett (2002) and Brett and colleagues (2004) suggested that the "Macedonia" of Ettensohn and others at this section was actually the Brannon equivalent (**figure 21**). This assessment is supported through recognition and correlation of the twin-seismite horizons with exposures at Donerail to the north. The twin deformed intervals are well-known in the upper Brannon (Donerail) interval. For more information and discussion of these intervals see below.

As defined, the Perryville Limestone is generally a tripartite unit with the lower Faulconer Bed transitioning from coarse-grained facies in the south into shaly nodular wackestone and packstone facies interbedded with calcarenite facies farther north in the Jessamine Dome as shown in outcrops and core from the Frankfort and Swallowfield regions. Throughout this region, McLaughlin and others (2004) show the Faulconer bed to be

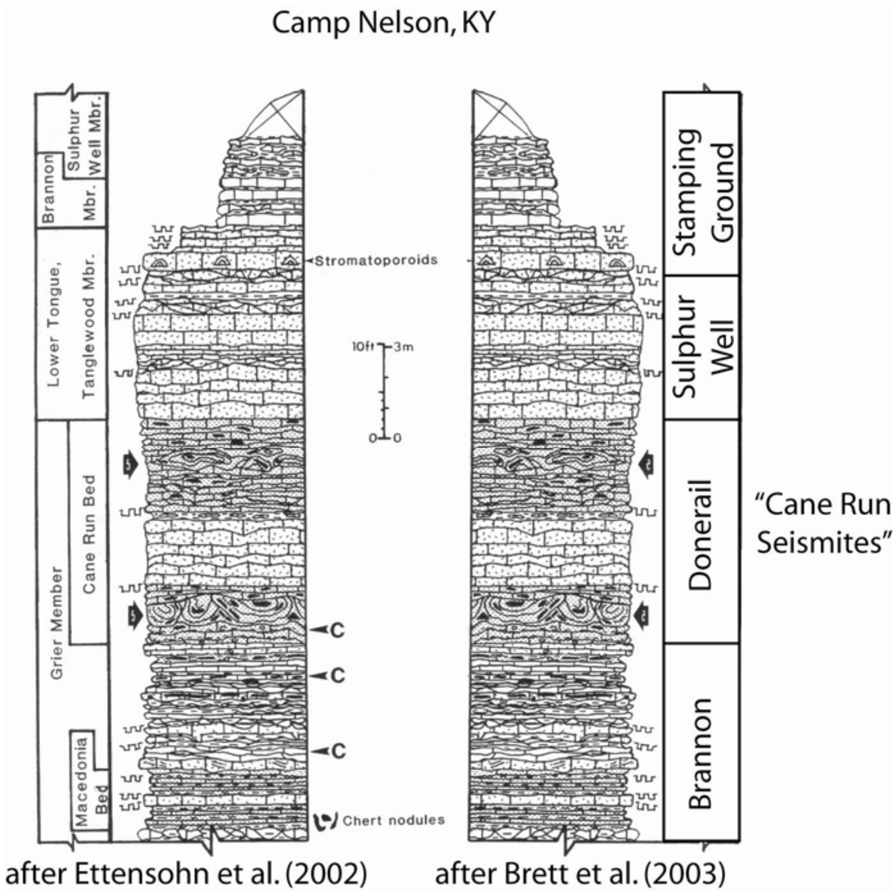


Figure 21: Comparison of stratigraphic terminology applied to the middle Lexington Limestone as observed in the Camp Nelson exposures on KY HWY 127 near the Kentucky River. Stratigraphic terminology on the left is after Ettensohn et al., 2002, and on the right is the terminology applied to the same unit based on correlations by McLaughlin and Brett (2002) and Brett et al, 2003. The primary discrepancy is differing recognition of the Macedonia Bed vs. Brannon. McLaughlin & Brett (2002) correlated this interval northward using the twin seismite horizons and found these two seismites to occur above the Cornishville stromatoporid unit and in the position of the upper Brannon deformed zones. The stromatoporoid biostrome unit at the top of the section is in turn correlated with the Stamping Ground interval.

approximately three meters in thickness and capped by a sharp surface at the base of the Salvisa. The overlying Salvisa Bed is dominated by fenestral micrite and calcilutite facies as mentioned, and ranges up to about five meters thick before it transitions northward into clean crinoid grainstone and shaly nodular wackestone to packstone facies of the Frankfort region and then fine laminated shale and limestone facies northward into the Sebree Trough. Without recognition of the Salvisa equivalent, the facies can be similar to the underlying Grier. Thus the unit above the coarse-grained Faulconer Bed at Frankfort was included in the Grier by Black and others (1965) and by Cressman (1973). Above the Salvisa, strata change sharply into thin-bedded coarse-grained calcarenites of the overlying Cornishville Bed. Both of the former units contain stromatoporoids. However the Cornishville shows evidence of significant storm disturbance and toppling of stromatoporoids and minor bioturbation. As shown in **figure 18** above, the Cornishville is capped by an interval of strongly mineralized hardgrounds at the base of the overlying Brannon. The Cornishville is the thinnest of the Perryville sub-units and approaches a maximum of three meters in thickness.

Similar to the underlying Grier, the diversity of fauna in the Perryville Member is high compared to lower Lexington. Bassler (1915) recognizes a large number of taxa from the Perryville Member – but does not distinguish between beds. Cressman (1973) did not supply a taxonomic list from the basal Perryville Faulconer Beds as the base was covered in most of his study areas. He did however indicate that the unit contained colonial corals (likely *Tetradium*), conspicuous mollusks, occasional brachiopods in some stringers, and stromatoporoids. Black shale partings in between Faulconer grainstones are also known to contain carbonized remains of green algae and occasional ostracods. McFarlan and White (1948) recognized the gastropod fauna as containing *Bellerophon troosti*, *Oxydiscus subacustus* and *Lophospira medialis*. Most of the latter are also known from the Bigby-Canon interval of Tennessee with which the Perryville correlates.

The Salvisa Bed has been studied in greater detail by the former author. Cressman (1973) lists a number of brachiopod forms including the characteristic *Hebertella frankfortensis*, and *Rhynchotrema cf. increbescens* (both of which are found in abundance in the Benson equivalent to the north), and at least three other genera of brachiopods. Most characteristic of the Salvisa Bed are the numerous bivalve forms observed – at least ten different forms have been identified. The appearance of these forms helps explain the more massive nature of the Salvisa interval due to more abundant bioturbation and more thorough mixing of beds. The Salvisa also contains three different ostracod taxa. McFarlan and White (1948) recognized three additional ostracod taxa and found these to be especially prevalent in the shalier portions of the unit rather than in the more massive limestones.

For the uppermost Perryville Cornishville Bed, Nosow and McFarlan (1960) specify the stromatoporoid *Stromatocerium pustulosum*, the brachiopods *Dinorthis ulrichi*, and *Strophomena vicina*, and the gum-drop trepostome bryozoan *Cyphotrypa frankfortensis*. Many of these are known from the upper Bigby interval of the Nashville Dome and eastward into the Valley and Ridge where they are also known from northwest Georgia (Wilson, 1949). A closely related sister species of *C. frankfortensis, C. pachymuralis,* is known from the slightly younger Coburn Formation of Pennsylvania and maybe a migrant from the south (Arens, 1988). Cressman (1973) added *Hebertella frankfortensis* (which appears to be ubiquitous in the Perryville), *Platystrophia colbiensis, Rafinesquina trentonensis,* and *Zygospira* sp. likely *Z. recurvirostris* to the roster of taxa. Kesling (1960) also reported the grainstone-rich Cornishville to contain at least one species of edrioasteroid from the hardgrounds near its top and just below the base of the Brannon. These edrioasteroids were classified as *Cincinnatidiscus carnensis* and compared to younger forms by Kesling (1960). If this classification is correct, this is the oldest occurrence of this genus, as most *Cincinnatidiscus* are known from the overlying Edenian and younger strata. Nonetheless, at least one of the younger *Cincinnatidiscus* forms (i.e. *Cincinnatidiscus stellatus)* has been reassigned to the genus *Cystaster* (Liberty, 1976). *C. carnensis* could potentially also be from this latter genus. If this is the case, its occurrence in the Perryville Member could place it within the age range of forms known from the Bobcaygeon and Verulam Formations of Canada. Overall, the faunas of the Perryville appear to represent a mix of early and late Trenton forms with a number of Black River-like forms returning for the first time since the lowest Curdsville.

Lexington Limestone: Brannon Member

The Brannon Member of the Lexington Limestone has been recognized as a distinct lithologic unit since work of Miller (1913) and before. Miller recognized the exceptional lithologic break between the Brannon and units above and below. Miller identified the Brannon as the lower part of the Flanagan Chert of previous workers. The lower part was characteristically argillaceous, fine-grained, even-bedded limestone, while the upper portion was often extremely contorted and contained an abundance of chert. The deformed beds were referred to as “flow rolls” (Nosow & McFarlan, 1960) but are considered to be “seismites” by most modern workers (Ettensohn, & Stewart, 2002, Ettensohn et al., 2002 a, Brett et al., 2004; etc.). In the Georgetown quadrangle north of Lexington and east of Frankfort, the Brannon ranges up to just less than five meters in total thickness. Further south, as much as nine meters have been reported (Cressman, 1973). Originally, the Brannon was the basal member of the Cynthiana Formation until Black and colleagues (1965) extended the top of the Lexington upward to include the Cynthiana Formation in its entirety. As described by Wahlman (1992) the Brannon is a medium gray, tabular-bedded calcisiltite unit interbedded with dark gray, calcareous shales (**figure 22**). In outcrop, the unit is generally sparsely fossiliferous, but the unit contains a few brachiopods and bryozoans. The Brannon is very similar to the Macedonia and the Logana Members below and has been characterized as a relatively deep-water, low-energy

Figure 22: Core photographs from the Swallowfield KGS Core C-203 (as shown previously) for the Brannon Member of the Lexington Limestone. The Brannon here is just over 5.3 meters thick and sharply overlies the Cornishville Bed of the Perryville and sharply underlies the Sulphur Well Member of the Lexington Limestone. The lower and upper Brannon are developed and there is some evidence for deformation in the upper zone of the Brannon (Donerail).

rhythmite facies (McLaughlin et al, 2004, Brett et al., 2004). In some cases, given these similarities it appears that the Brannon interval has been mistaken for the Macedonia and vice versa.

The base of the Brannon is extremely sharp and easy to recognize in outcrop and cores where the shaly Brannon often weathers in recess relative to the underlying and overlying units (**see figure 19**). The characteristic mineralized hardgrounds at the top of the underlying coarse-grained Cornishville Bed are easily recognized. This contact is usually between 55 and 60 meters above the base of the Lexington Limestone (Cressman, 1973). As shown in **figure 22**, the contact is often very sharp and planar or occasionally gradational through an interval of nodular beds. Recognition of a K-bentonite in the Brannon (by Black et al., 1965) at several outcrop localities and in cores helps indicate that the Brannon itself is a nearly synchronous unit across its exposure area. This K-bentonite is the Quisenberry Road K-bentonite after Brett and colleagues (2004) and used herein.

As suggested by Black and others (1965), the top of the Brannon is also easy to recognize beneath the coarser limestone ledges of the overlying unit (their Tanglewood) where well-exposed. Cressman (1973) referred to this tongue of the Tanglewood as the Sulphur Well Member of the Lexington Limestone. The Sulphur Well Member is a massive, rubbly-weathering, irregularly-bedded, bryozoan-rich limestone and coarse-grained calcarenite unit that becomes cross-bedded especially in its upper part. The deformed beds of the upper Brannon (referred to as the Donerail by Brett et al, 2002) are shown to be slumped downward into the underlying shalier lower Brannon and in some cases, the overlying Sulphur Well interval also demonstrates similar deformation – suggesting that at least in some regions as many as three deformed horizons can be observed in the Brannon to lower Sulphur Well. These coarser-grained upper beds and the deformed strata have been intensely studied by previous authors (McLaughlin, 2002, Ettensohn et al., 2002)

Given the shaly-nodular nature of the Brannon, this interval is easily identified in outcrop and core. On wireline logs this shaly zone can be recognized in the middle of the Lexington where it is most pronounced and has not been truncated by overlying units. Cressman indicated that in the southern Jessamine Dome, the basal Sulphur Well contains pebble intraclasts of Brannon calcisilts indicating that the upper contact of the Brannon contact is likely unconformable. The Brannon was thought to pinch out below the overlying Sulphur Well in most areas outside of the central Jessamine Dome (Black et al., 1965). For instance, the unit to the west of the Kentucky River just west of Cornishville is thought to be significantly thinned or truncated from above. These authors also suggested that the Perryville did not extend north of the Georgetown quadrangle where it was recognized by Miller (1913). Moreover, they also indicated that the Brannon was not present east of Lexington and north of Millersburg and Winchester in Clark County. Thus, it was suggested that the Brannon was deposited in a centralized area centered in the Lexington region east of the Kentucky River.

Nonetheless recent work has shown that the Brannon, although substantially thinner is present in the vicinity of Frankfort and the northern Georgetown quadrangle. It rapidly expands northward where it is known from cores in the Swallowfield area north of Frankfort through the northern Kentucky to southern Cincinnati region (**see figure 22 & 23**). In this northernmost area, the Brannon correlates into the Sebree Trough and merges with the larger shale-dominated succession of the "Utica" of former workers (Bergström and Mitchell, 1986).

Faunally, the Brannon as mentioned is relatively barren but has a bryozoan-dominated fauna which preferred the deeper-water siliciclastic-rich environment. Cressman (1973) lists a group of at least nine different bryozoans and several species of brachiopods that appear to be holdovers from the underlying Perryville. McFarlan and White (1948) identified this interval as especially characterized by the bryozoan *Crepipora spatiosa* (now *Papillalunaria spatiosa* Utgaard, 1968), *Eridotrypa aedilis*, *Peronopora milleri* and a large massive bryozoan referred to as *Heterotrypa* sp. The former cystoporate bryozoan has been compared to forms (*Papillalunaria perampla*) recorded from the Decorah interval in Minnesota near the Chatfield type-region but appears transitional to younger forms. Of course as with underlying units the Brannon continues to have abundant *Prasopora falesi* (now considered *P. simulatrix*). Also recognized in the Brannon are specimens of *Brachiospongia digitata*.

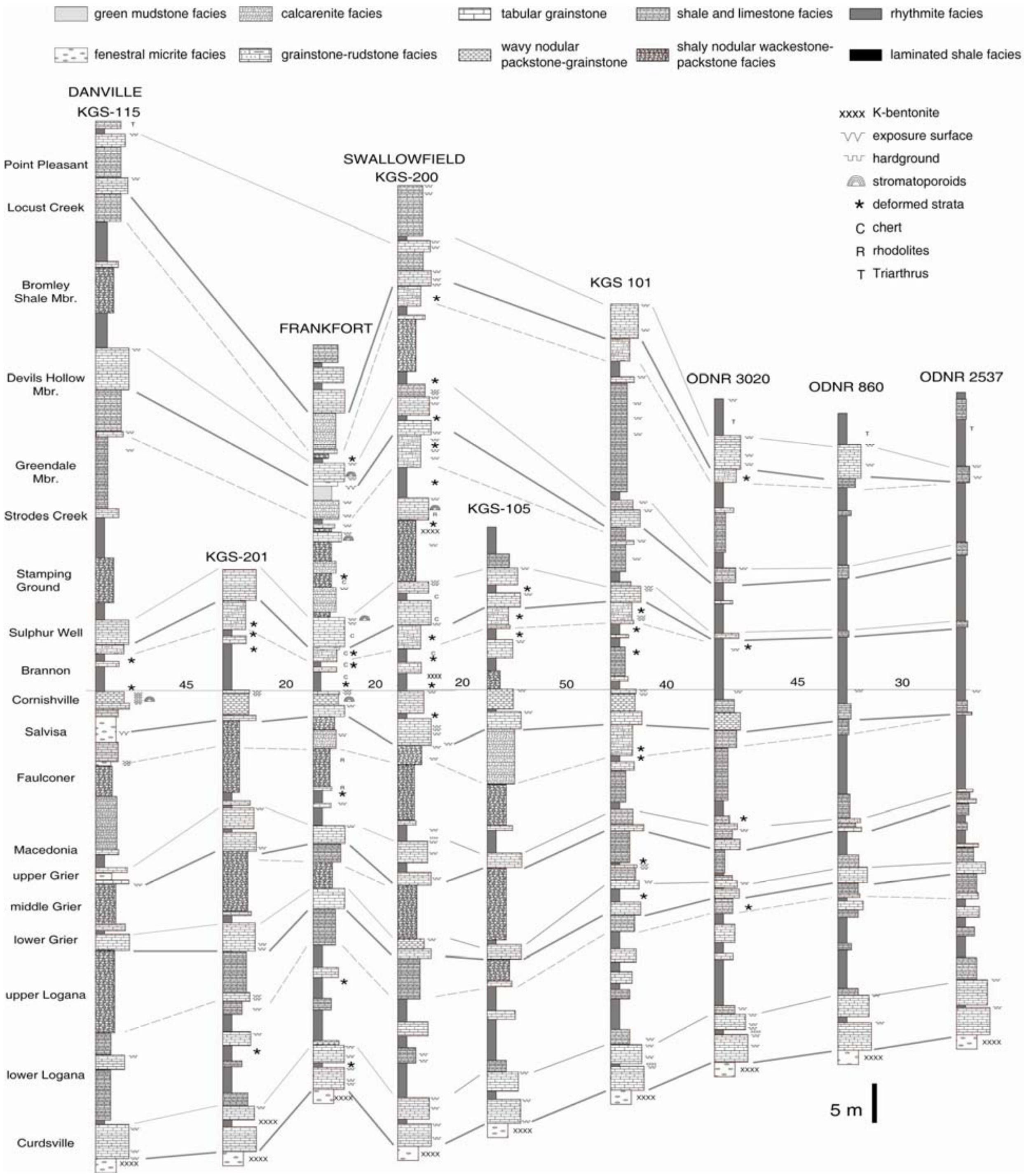


Figure 23: Stratigraphic cross-section for the Lexington Limestone interval from the region of Danville northward along the Cincinnati Arch to Butler County Ohio (total ~270 km). Datum plane is established at the base of the Brannon Member of the Lexington Limestone – which forms the base of the Cynthiana Formation of previous workers. Stratigraphic section modified after McLaughlin et al., 2004.

Lexington Limestone: Sulphur Well Member

The Sulphur Well Member of the Lexington Limestone was first defined as a member of the Cynthiana Formation by McFarlan (1943), and later incorporated into the expanded Lexington Limestone (Black et al., 1965). The Sulphur Well was also referred to previously as the Woodburn phosphatic member (Miller, 1913; McFarlan & White, 1948; Nosow & McFarlan, 1960). Described by these early workers as a phosphate-rich crystalline limestone, it was recognized as especially important due to its fossil content. Cressman (1973) described the unit as a poorly-sorted bryozoan rudstone facies with minor calcisiltites in irregular to lenticular beds that are separated by thin shale partings. The facies is considered a lateral equivalent of the upper lower Tanglewood of Ettensohn (2002). This description is not entirely representative of the original concept of the Sulphur Well as a crystalline limestone, but is likely a correlative facies of this unit. McLaughlin and colleagues (2004) and McLaughlin (2007), described the unit as typically developed into a well-bedded calcarenite facies (with numerous bryozoans) in the area north of Danville northward through the Frankfort and Swallowfield regions before it begins to grade into the nodular facies described by Cressman (**see figure 22**). Cressman indicated that the Sulphur Well resembled a Grier-like lithology; however, this does not seem to be the case in much of the outcrop and subcrop region.

The Sulphur Well was considered to range between three and ten meters in total thickness in its exposure region from south to north respectively (Cressman, 1973). This author suggested that the Sulphur Well thinned to the north due to facies change into the base of the "Clays Ferry Formation" (base of the Stamping Ground Member). In the Swallowfield C-203 core (as shown in **figure 22**) the Sulphur Well is just over six meters in thickness. As shown in **figure 23**, McLaughlin and colleagues (2004) show this to be the case as the Sulphur Well interval shows evidence of condensation into a thin unit northward into Ohio.

Cressman described both the basal and upper contacts of the Sulphur Well to be sharp and relatively planar in the central Jessamine Dome. The basal contact is usually distinct as coarser grainstones (bryozoan-rich calcarenites) sit above shaly-nodular calcisiltites of the underlying Brannon. In many localities, especially south of Frankfort and at Camp Nelson, the basal contact sits above the Donerail seismite interval discussed previously (see McLaughlin et al., 2004). The coarse-grained calcarenites of the Sulphur Well are sharply overlain by the shaly nodular wackestone to packstone lithologies of the Stamping Ground Member. To the east, the top contact of the Sulphur Well was considered to be more gradational as the unit graded from bryozoan-rich limestone into bryozoan-rich shales at the base of the Millersburg Member (base Clays Ferry Formation of older workers).

The Sulphur Well interval is a critical unit in terms of biostratigraphy of the Lexington Limestone. This is because the Sulphur Well records the incursion of some new taxa not previously recorded in the region. The Sulphur Well faunas and those of the remainder of the Lexington are included in Ordovician time slice 5b of Webby and colleagues (2004). In addition to recording the return of a number of coral taxa that had been absent from the region for some time (i.e. *Columnaria halli*, and *C. alveolata*), the Sulphur Well was also documented to contain *Constellaria teres* (Nosow & McFarlan; 1960). This particular form of cystoporate bryozoan has not been recorded from any lower stratigraphic intervals and appears to be coincident with the appearance of this form elsewhere in the GACB. In addition to this bryozoan form, Cressman (1973) includes at least thirteen different bryozoan taxa. Six of these forms are not found in immediately underlying strata, but are found lower in the section (typically from the lowest

Lexington strata). Others appear about this time and extend upward into overlying strata. This part of the Lexington shows the rapid evolution and diversification of several forms of bryozoans including the *Peronopora*. Pachut and Anstey (2002) suggested that a major event allowed for all species of *Peronopora* (eight major types based on cladistic analysis) to appear within the Lexington Limestone by the top of the Millersburg Member. Numerous forms found in the Sulphur Well and overlying units suggests a nearly simultaneous first appearance and is thought to represent a punctuated event or series of events, which opened a new taphonomic or paleoenvironmental window during this time (Pachut and Anstey, 2002).

Cressman (1973) also identifies the typical Lexington assemblage of brachiopods including *Hebertella frankfortensis, Rafinesquina* sp., and *Zygospira* sp.. Nosow & McFarlan also identified additional brachiopods including *Platystrophia colbiensis* and *Rhynchotrema increbescens*. Interestingly, these authors also noted the rather large numbers of the diminutive gastropod *Cyclora minuta* (now considered *Cyclonema minuta*). This particular gastropod was known from occurrences in Nashville Dome region first in the Curdsville (lower Hermitage) of Tennessee, and then from the Bigby of the same region (Wilson, 1949). It makes its first appearance in the Brannon, but becomes very abundant on some bedding planes in the Sulphur Well (the Woodburn of Nosow & McFarlan, 1960). This taxon becomes abundant in Maysvillian to Richmondian strata, but its appearance in the Brannon to Sulphur Well interval may be correlative with an incursion of these forms in the Martinsburg Formation of the Cumberland Valley in Pennsylvania and Maryland as reported by Bassler (1915). In the latter region, they often occur in association with *Corynoides americanus* graptolites.

Also interesting in this interval in Kentucky, is the report of *Rafinesquina* sp. by numerous authors. The form found beginning in this level is not the characteristic *R.*

trentonensis form found lower in the section, but it is also not entirely characteristic of forms found in the overlying Cincinnatian. The forms found in the Sulphur Well and in the Millersburg are generally thought to be *R. winchesterensis* or slight variations of *R. alternata* a form known from the Upper Trenton and younger strata. Thus again, this particular assemblage appears to represent a transition between older and younger faunas, and was likely influenced by a significant and important paleoenvironmental shift recorded across a wide portion of the GACB.

Lexington Limestone: Stamping Ground Member

Cressman (1973) was the first to recognize the Stamping Ground Member of the Lexington Limestone for the fossiliferous limestone and shale interval immediately overlying the Sulphur Well. It was named for the town of the same name in Scott County, Kentucky. The unit is typically reported to be about three meters thick but ranges upward to nearly seven meters in total thickness. Cressman used the term Stamping Ground for the strata separating two tongues of the coarse-grained Tanglewood Member of earlier authors. As it was quite nodular and shaly, it was considered a similar lithology to the Millersburg and the Greendale, but viewed as non-continuous with either of those units. Typically, the Stamping Ground is a light- to medium-gray bioclastic limestone often with a calcisiltite matrix. The interbedded shales of the unit are olive gray in color. McLaughlin and colleagues (2004) described the Stamping Ground as nodular shaly packstone passing upward into wavy nodular- to cross-bedded packstone to grainstone unit often containing abundant overturned and fragmented stromatoporoids and occasional *Solenopora*.

The Stamping Ground sits sharply above the underlying Sulphur Well calcarenites (a tongue of the Tanglewood Limestone). Prominent hardgrounds are developed at the contact and immediately below the unit as recognized by McLaughlin and colleagues (2004) and McLaughlin (2007). In some localities the contact shows development of bryozoan-algal bioherms that are draped by shales and argillaceous limestones of the Stamping Ground. These prominent hardgrounds appear to parallel those at the base of the underlying Brannon and can be used as excellent synchronous marker horizons. Nonetheless, it was clear from his description that Cressman (1973) viewed the Stamping Ground as a lenticular body that was (similar to underlying units) restricted to a region within the central Jessamine Dome area and one that graded laterally to the south and north into coarse-grained calcarenite facies of the Tanglewood. Moreover, eastward to the region of Bourbon County, the loss of the distinctive Sulphur Well calcarenites and the Brannon prevented separation of the Stamping Ground from the underlying Grier in this region (Cressman, 1973). This appears somewhat in contrast to the notions of Black and Cuppels (1973) who suggested that a Tanglewood calcarenite unit was indeed recognizable between the Grier and the base of their Millersburg Member – a Stamping Ground equivalent (**figure 24**). Nonetheless as shown by McLaughlin and others (2004), the Stamping Ground interval, despite some lateral facies change, is recognizable and correlatable from the Danville region (Lawrenceburg Core KGS C-115) northward to the Ohio border where it passes out of nodular facies and into tabular rhythmite facies south of the margin of the Sebree Trough (**see figure 23**). Eastward, this particular facies is also developed and referred to as a tongue of the Clays Ferry (within the Millersburg Member) by Black and Cuppels (1973).

The upper contact of the Stamping Ground was not discussed explicitly by Cressman (1973). Nonetheless in his type section description of the Stamping Ground, he did indicate that

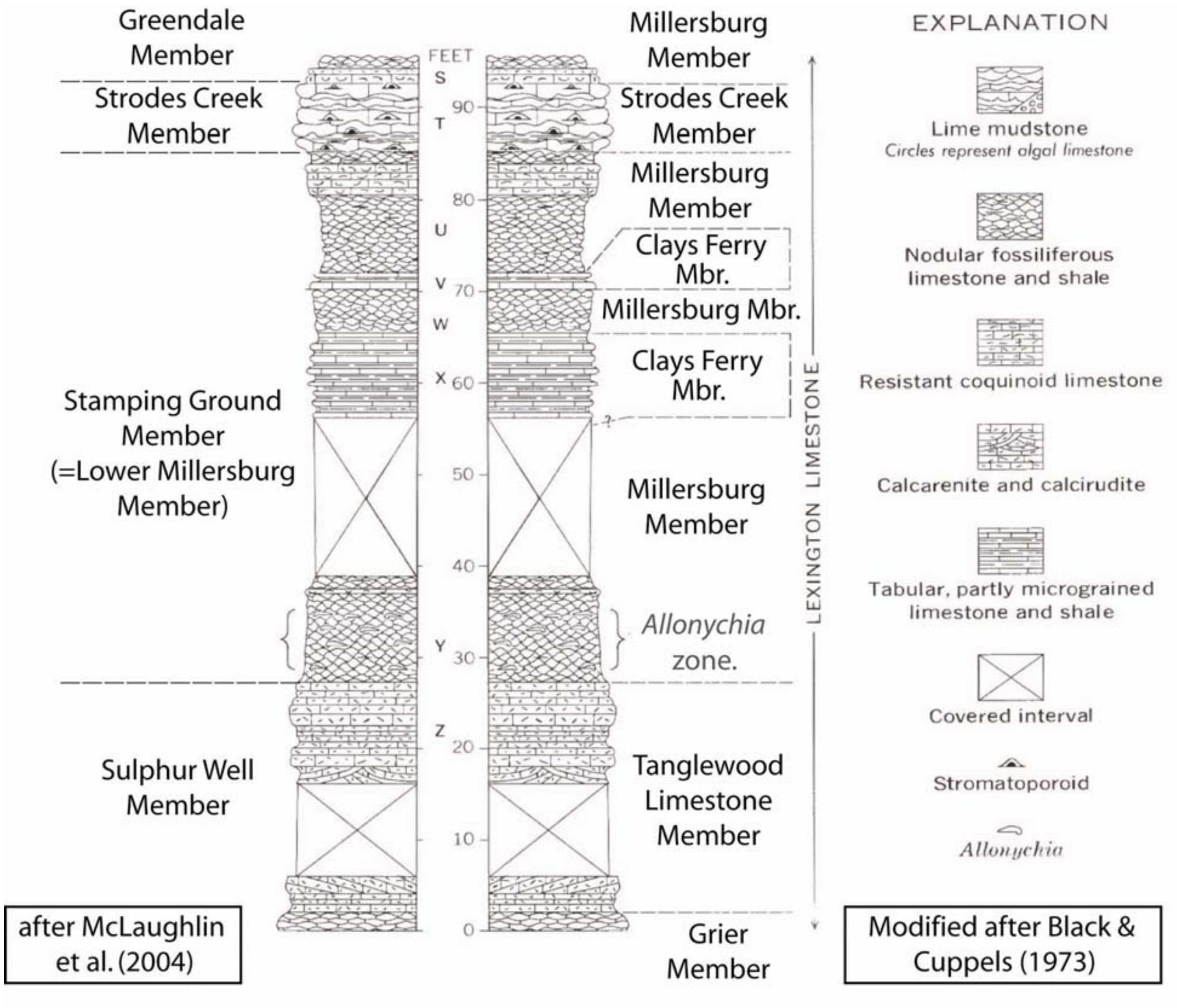


Figure 24: Stratigraphic context of the Millersburg Member of the Lexington Limestone and the position of the Strodes Creek Member of the Lexington Limestone in the type region of the Strodes Creek (Clark County, Kentucky) after Black & Cuppels, 1973. Also shown are the lithostratigraphic terms recognized and used by McLaughlin and colleagues (2004) for the equivalent interval along the central axis of the Cincinnati Arch. It is suggested that the base of the Stamping Ground Member is equivalent to the base of the Millersburg Member, although

the Stamping Ground was overlain by light-gray, fine- to medium-grained, bioclastic calcarenites or grainstones. This unit was referred to as yet another tongue of the Tanglewood Member, but was not as thick as other calcarenite units. This capping calcarenite was described to contain a faunal assemblage dominated by large trilobites, brachiopods, algae, and numerous large stromatoporoids which is somewhat unlike other Tanglewood faunas. Based on correlations by McLaughlin and colleagues (2004), this upper calcarenite is likely the Strodes Creek equivalent. Slightly below this bed, is a K-bentonite that has been recognized in outcrops and core especially in the area immediately north of Frankfort, Kentucky. McLaughlin and colleagues and McLaughlin (2007) refer to this K-bentonite in the upper Stamping Ground as the Swallowfield K-bentonite based on its development in that region. Similar recessive-weathering re-entrants have been noted at several outcrops below the Strodes Creek level and likely represents this same K-bentonite.

In contrast to underlying units, the Stamping Ground is dominated by taxa including larger brachiopods (*Rafinesquina* and *Platystrophia*), and numerous gastropods and mollusks (although it contains some bryozoans including *Constellaria teres* and a form of *Peronopora*). Although not identified in the faunal lists of Cressman (1973), the epifaunal byssate bivalve *Allonychia flanaganensis* was reported from the equivalent strata in the Clark County region (Black and Cuppels, 1973). The unit also contains recognizable graptolites (*Diplograptus* sp.) which were also reported from the underlying Brannon (Beecher, 1889; Cressman, 1973). Little work has been done to identify the specific form of graptolites found in these beds so it is not clear if they have any value in correlation with the graptolite biostratigraphy in the GACB region at this time, but would be worth further investigation. Cressman's list of gastropods included *Cyclora* sp., *Loxoplocus* (*Lophospira*) sp., and *Microceras* sp.. *Cyclora* sp., is thought to be the juvenile form of *Cyclonema* (McFarlan & White, 1948 report *C. varicosum* from this interval) and *Microceras* sp. is now considered to belong to the genus *Cyrtolites.* Many of these gastropod forms are characteristic of the upper Lexington and overlying Kope Formation and are less characteristic of underlying units and appear coincident with the significant and dramatic return of coralgal facies.

Taha and colleagues (2001) showed that the Stamping Ground also contained a faunal association that included *Solenopora* (suspected red algae), *Tetradium*, and some ramose and sheet-like bryozoans. In addition these authors identify the occurrence of the stromatoporoid *Labechia huronensis* which may occur in biostromal masses at the caps of small, meter-scale cycles. These are best developed in coarser "Tanglewood-Stamping Ground" facies near Frankfort and Lexington, but are found in shalier-nodular facies elsewhere – where they are usually smaller. However, they appear in successively higher cycles above the top of the Sulphur Well. As suggested by these authors, the restriction of *Labechia* to the Stamping Ground interval appears to be dependent on a higher order control that likely was related to warming.

Lexington Limestone: Strodes Creek Member

The Strodes Creek Member was first described by Black and Cuppels (1973) from exposures on the eastern side of the Jessamine Dome (Strodes Creek, Clark County, Kentucky). It was considered a member of the Lexington Limestone and was well-exposed in road cuts along US 227 and interstate I-64. The Strodes Creek was defined as a unit dominated by grayish-brown to gray limestone interbedded with minor amounts of reddish-brown to gray, argillaceous calcisiltites and somewhat fissile shales. The Strodes Creek, especially in the Frankfort, KY region and elsewhere contains easily recognized ball-and-pillow structures (Black & Cuppels, 1973), and numerous stromatoporoids and algal (rhodolite?) nodules. The Strodes Creek is usually a thin interval, most commonly ranging up to two meters in total thickness. As the seismites and algae conglomerates are not always well-developed, it has been overlooked in many sections and is often considered to be a lens or tongue of the Tanglewood within the Millersburg (Black & Cuppels, 1973).

Despite its classification as a lens or tongue, McLaughlin and colleagues (2004) were able to recognize the Strodes Creek over a wider area in outcrop and core to the west and north of the type region recognized by Black and Cuppels (1973). It is most often dominated by medium dark gray highly fossiliferous, wackestone facies, but can be developed into coarser-grained, nodular wacke- to packstone and calcarenite facies containing abundant stromatoporoids especially in the Lexington to Frankfort area. It does, however, typically contain a fauna rich in rhynchonellid brachiopods and contains more bryozoans than the underlying Stamping Ground. North and south of this area, the unit transitions into shaly nodular and even rhythmite facies especially in the Cincinnati area and the Cumberland Saddle respectively (McLaughlin 2007).

The base of the Strodes Creek is drawn at the top of the Stamping Ground (or top of the lower tongue of the Millersburg Member of the Lexington of Black & Cuppels, 1973). The contact appears to be sharp but conformable with the base of the Strodes Creek developed as the cap of a shallowing-upward cycle out of the underlying shaly-nodular Stamping Ground. At this level, nodular limestones grade upward into coquinal packstones and minor grainstones and suggest a shallower clean skeletal grainstone-rudstone facies. The top contact of the Strodes Creek is often heavily stained and mineralized through the weathering of pyrite to limonite. This hardground contact is immediately overlain by another interval of nodular to planar fine-grained rhythmite facies equivalent to the Greendale or upper tongue of the Millersburg Member (of Black and Cuppels, 1973). As mentioned, the Strodes Creek also contains an interval of well-developed ball and pillow structures that aids in correlation of this unit only locally. It is not consistently recognized.

The Strodes Creek contains a diverse assemblage of fossils. Black and Cuppels (1973) recognize a number of corals, brachiopods, a diverse assemblage of gastropods, bivalves, and ostracods (as many as 15 different forms), as well as a number of bryozoans. Also included is the conodont *Rhipidognathus symmetrica* – a form that is known well up into the Richmondian. The bivalves recognized by the former authors are significant in that they have typically been a minor to absent component of much of the Lexington before this level. The occurrence of both infaunal (*Modiolopsis* aff. *M. simulatrix*) and epifaunal forms including *Ambonychia* are important in that these taxa become significant components of later faunas in the region – and are generally absent from the lower Lexington – although some forms have been noted in the Logana and the Salvisa as discussed previously. Important from the list of gastropods includes the bellerophont *Buchania* and the pleurotomarid *Murchisonia* (*Lophospira*) *milleri*. Both of these forms are important as they are also noted from the upper Trenton of New York where they were collected by Walcott.

Frey (1992) also noted the interval to be especially important in terms of nautiloids diversity. His work indicated that the lower Millersburg Member (Stamping Ground) contained only the orthocerid *Isorthoceras albersi*. In contrast, the Strodes Creek from Winchester, Kentucky shows the return of a rather large orthocone referred to *Polygrammoceras*? cf. sp. *A*. Frey also noted specimens of a small cyrtocone that he considered was similar to *Faberoceras sonnenbergi*. Combined with an assemblage of six additional taxa recognized by Flower (1942) from what Frey considered to be a lateral equivalent of the Strodes Creek at Cynthiana, the nautiloid fauna of the Strodes Creek shows the first evidence of having recovered significantly from the earlier extinction. Most of the forms recognized by Flower include the slender cyrtocone *Oncoceras*, specimens of the orthocerid *Isorthoceras* as well as the actinocerids *Deiroceras* and *Orthonybyoceras* and several others. In Frey's estimation based on similarities in these faunas between Winchester and Cynthiana, and the style of their preservation and associated lithologies, these faunas are likely coeval. Thus the assemblage, as he states, is the most diverse nautiloid fauna in the Lexington Limestone. Although the fauna contains a number of endemic taxa, it is considered to be most similar to forms found in the Catheys Formation of central Tennessee. This group of nautiloids seems to persist for some time and subsequently gives rise to forms found in the much younger Leipers Limestone (Frey, 1995). Clearly the Sulphur Well, Stamping Ground and Strodes Creek succession represents a significant faunal and paleoenvironmental shift in the Jessamine Dome region at this time. This diversification event is also noted in the other time equivalent strata including in the Coburn Formation of central Pennsylvania, the Rust Formation of New York State (Frey, 1995), and evidently in the Catheys Formation of the Nashville region (Bassler, 1932).

Lexington Limestone: Greendale Member

Foerste (1906) first recognized the Greendale as an argillaceous limestone and calcareous shale unit in the Cynthiana Formation that was underlain by the Lexington Limestone and overlain by the Point Pleasant Bed of the Cynthiana (Nicholas Limestone of central Kentucky; McFarlan & White, 1948). The Greendale was not consistently applied by most workers and McFarlan and White (1948) suggested that the Greendale was at least in part a lateral equivalent of the Millersburg Limestone. Cressman (1973) identified the Greendale and re-designated it as a lentil within the Lexington Limestone based on the assumption that it occurred in a small region just north of Lexington and that it was entirely surrounded by laterally equivalent calcarenites of the Tanglewood Limestone. In the type section, he described the unit as a three meter –thick, fossiliferous limestone and shale. Cressman identifies the Greendale as a fine-grained olive-gray to light-gray calcisiltite unit that is typically very nodular and occasionally interbedded with thin shale beds and irregular beds of coarse-grained calcarenites. It thickens slightly in a nearby core to just over four meters, but is often truncated below the overlying calcarenites of the Devils Hollow Member of the Lexington Limestone.

Cressman correlated the unit eastward a short distance where just over one meter is recognizable in sections east of Lexington – but included this lens in the Tanglewood because it is so thin. Further to the south and west, another shaly limestone unit was also recognized – but considered a tongue of the Millersburg – and not an equivalent of the Greendale. Nonetheless, work by McLaughlin and colleagues (2004) recognize the Greendale as an important stratigraphic interval immediately below the Devils Hollow calcarenites. As indicated by McLaughlin (2007), the Greendale is traceable across central Kentucky by the presence of the heavily mineralized hardground at its base with the underlying Strodes Creek Member. In addition, recognition of a calcisiltite-dominated deformed interval is also more or less easy to identify in outcrops especially as this particular horizon is often found to contain an epibole of *Hindia* sponges.

The Greendale is especially important as it appears that its upper contact is unconformable and shows some evidence for erosional truncation at the base of the Devils Hollow. The contact between the Greendale and the lower Devils Hollow is typically flat and sharp; however, near Frankfort this contact is demonstrated to have moderate relief (McLaughlin, 2007). McLaughlin and colleagues (2004) correlate the Greendale section south of the Frankfort, Kentucky area (where it is especially thin) and recognize the contact above the deformed interval of the Strodes Creek. In the region of Danville, the Greendale becomes more rhythmically bedded and returns to its calcisiltite-dominated facies below the level of the Devils Hollow (**see figure 23**). Aside from the *Hindia*-type sponges, a faunal list has not been published for this particular stratigraphic interval.

Lexington Limestone: Devils Hollow Member

McFarlan and White (1948) named the Devils Hollow Member of the Lexington Limestone for the interval immediately above the Greendale. At its type locality and in the immediate vicinity, Cressman (1973) defined the unit as about 4.5 meters of porous, coarsely crystalline massive limestone containing abundant gastropod shells. This was overlain by an interval of three meters of compact ostracod-bearing calcilutite similar to the Salvisa Bed of the Perryville and the Tyrone Formation of the High Bridge Group. In total, the Devils Hollow is reported to range from 7.5 meters to just over nine. In the interval near the contact of these two lithologies both facies inter-tongue (become cyclically-bedded) and are easily mapped. Thus the Devils Hollow is a two-part unit with a coarse, lower unit and an upper finer-grained, shallower water unit. Within the lower to middle Devils Hollow sub-unit the facies occasionally contain an interval of greenish mudstone facies marked by desiccation-cracks. This facies grades into calcarenite facies to the north and south. It commonly shows cross-lamination with mud drapes and herringbone cross-stratification interpreted as tidally influenced (McLaughlin et al., 2004). Collectively this facies records a significant jump to the north of the shallowest water facies – likely tied to fault reactivation producing uplift and slight subsidence in different areas of the Lexington Platform.

The Devils Hollow rests immediately above the Greendale lentil of the Lexington, except for in areas where the Greendale is truncated. In the latter case, the Devils Hollow appears to rest on lower beds of Tanglewood Limestone (the Strodes Creek). In the Frankfort region, the Devils Hollow again lies above a bryozoan-rich shale and limestone unit containing abundant

Constellaria, which is correlated with Millersburg and is likely an equivalent of the Greendale Lentil. Immediately above the Devil's Hollow is a facies similar to the Millersburg (and is considered as the uppermost Millersburg by Cressman, 1973). In contrast, McLaughlin and colleagues (2004) identify this interval as the Bromley Shale equivalent of northern Kentucky.

The Devils Hollow interval is a distinct lithologic unit, is composed of multiple small meter-scale parasequences, and is easily correlated. It too has been documented to have a deformed horizon. Both north of Frankfort and in the subsurface of the Cincinnati region, the lower calcarenite facies shows extensive deformation especially in the area of Swallowfield. In some cases, the deformed beds appear to be developed in channel-forms oriented roughly north-south (McLaughlin et al., 2004). As mentioned the lower portion of the Devils Hollow is also locally composed dominantly of cemented gastropod coquinas but maybe developed into brachiopod pavements. In many places, these coquinas weather as chert-rich intervals and might be considered as unsorted biosparrudites (Cressman, 1973). There is typically evidence of significant reworking of fossil materials in some beds.

The upper Devils Hollow (or "fine-grained" phase of Cressman, 1973), shows excellent cyclic bedding between beds of greenish-gray to light-gray calcilutites. This interval is dominated by a diminutive ostracod and gastropod assemblage. These commonly occur where laminae become slightly irregular and crinkly and suggest evidence for cyanobacterial mats. The association of ostracods and grazing gastropods suggests that this facies was somewhat restricted. Associated with these laminated beds that often cap small-scale cycles, are a number of brownish-gray, burrow mottled beds that are slightly coarser in nature and have fragments of crinoids, bryozoans, and other materials. The uppermost contact of the Devils Hollow is recognized by the close stacking of as many as four different hardground surfaces that show evidence for sediment starvation and mineralization just below the base of the Bromley Shale and its characteristic rhythmite facies (McLaughlin et al., 2004).

In contrast to the Strodes Creek interval below, the Devils Hollow complex is again distinctly less-diverse. The unit is reported by Cressman (1973) to contain *Zygospira*, three different species of gastropods and bellerophonts, a bivalve referred to as *Colpomya constricta* as well as at least two species of ostracods including the excellently preserved *Teichochilina jonesi*. Other fossil material is noted, but taxonomic identification is difficult due to taphonomic conditions. Most of these forms appear either to be endemic or in the case of the at least one of the gastropods (long-ranging). These beds also contain a stromatoporoids although they are not as large or as abundant as they are in underlying units (Cressman, 1973). Cressman suggested the Devils Hollow complex represented deposition in a northwest-trending lagoon that opened to the southwest into more open marine associations.

Lexington Limestone: Bromley Shale Member

Bassler (1906) first identified the Bromley Shale in the vicinity of the Ohio River in northern Kentucky. It was first described as an interval of drab to dark blue shales outcropping in the bank of the Ohio River immediately south of Cincinnati. In total they were measured to about nine meters in thickness, contained abundant *Dalmanella* and trilobite remains, and occurred just below the interval referred to as the Point Pleasant Formation. Luft (1971) included the Bromley as a member of the Point Pleasant Formation during his mapping in northern Kentucky, and Weir and colleagues (1984) in turn modified it as the Bromley Shale Bed of the Point Pleasant Tongue of their Clays Ferry Formation. This was the basal Clays Ferry interval of Weir and Green (1965), and is the uppermost portion of the Millersburg Member of the Lexington Limestone (of Cressman, 1973).

As reported by McLaughlin and colleagues (2004), the Bromley shows some lateral facies change from south to north near its type section where it is typically a calcisiltite and shale rhythmite facies. In the central Jessamine Dome area around Frankfort, the Bromley is slightly more nodular and interbedded with coarser calcarenite beds and is distinctly less shaly. In this area it typically is just less than ten meters in thickness. From Frankfort, the Bromley shale rapidly expands both north and south to as much as twenty five meters and becomes significantly more shaly although it is still somewhat nodular especially in the area of Danville and Swallowfield. North of Swallowfield into northern Kentucky and southwestern Ohio, the Bromley is developed into the uniformly-bedded fissile calcisiltite and shale facies typical of the outcrop exposures in the Ohio River Valley near Cincinnati. As reported by Cressman (1973) the lower part of the Clays Ferry (the Bromley equivalent) consists of medium- to dark-gray argillaceous calcisiltite interbedded with brachiopod rudstones and crinoid calcarenites. The calcisiltites are often relatively barren but may contain crinoid columnals and gastropod packstones. Typically the brachiopod packstones to rudstone beds contain *Rafinesquina*, and or *Dalmanella* – the former of which can often be observed to be stacked vertically and shingled.

As mentioned the lower contact of the Bromley is placed at the top of the uppermost of the well-developed hardground surfaces on top of the underlying Devils Hollow. As suggested by McLaughlin and others (2004), the Bromley is developed into several stacked small-scale cycles that gradually shallow-upward. However, the uppermost Bromley shows evidence of another sharp and often channeled erosion surface between the deep-water rhythmite facies of the lower Bromley Member and the shallow water calcarenite facies of the upper Bromley

(Locust Creek sub-unit, after McLaughlin and Brett, 2007). In northern Kentucky, the dark organic rich shales and calcisiltites of the Bromley form the lowest part of the section, and grade upward into the Locust Creek sub-unit that is represented by an interval of deformed argillaceous calcarenites (**see Figure 23**). The details of the Locust Creek deformed interval have been reported by McLaughlin (2002). This succession is thus sharply overlain by the massive coarse-grainstones of the overlying Point Pleasant Formation (after McLaughlin et al., 2004).

After the relatively depauperate fauna found in the Devils Hollow, the Clays Ferry – Bromley interval is again diverse and fossiliferous. Cressman (1973) reports a fauna that contains over twenty different bryozoan taxa, more than a dozen brachiopod taxa including forms typical of the upper Trenton of New York, and the Kope Formation. A diverse group of gastropods and as many as eight or more Trilobite taxa are recognized including a reappearance of *Cryptolithus* and other forms that are also found in the Rust Quarry and Prospect Quarry members of the uppermost Trenton in New York. Correlation of the Bromley northward into the subsurface of Hamilton and Butler Counties shows that the lower (dark shale-dominated sub-unit) Bromley to also contain the trilobite *Triarthrus becki* and several other taxa including the distinct crinoid *Merocrinus*. These forms appear to represent the first incursion of dysoxic biotas into the Cincinnati Arch through the Sebree Trough and may have been delivered via new deep-water connections to the east where such faunas had been present for some time.

Lexington Limestone: Point Pleasant Member

Orton (1873) was the first to recognize the Point Pleasant in Clermont County, Ohio along the northern bank of the Ohio River. It was defined as an interval of about fifteen meters of massive blue limestone interbedded with concretion-bearing shales. The unit was considered the lowest of the Cincinnati Group. The top of the Point Pleasant was demarcated at the base of the Fulton sub-member of the overlying Kope Formation. In mapping activities on the western side of the Cincinnati Arch in northern Kentucky, Swadley (1975) considered the Point Pleasant to be a tongue of the Clays Ferry Formation (following the work of Weiss et al., 1965). In this area it was described as containing two dominant lithologies: a medium- to coarse-grained bioclastic limestone often in thick planar to wavy beds, and thin, interbedded micro-grained limestones that were sparsely fossiliferous. Silty shales were also important components and ranged up to fifty centimeters in thickness. In most of the western region, the base of the unit was not exposed and a total thickness was not established in this region. Cressman (1973), and Wahlman (1992) included a much thicker interval in the Point Pleasant than assigned by previous workers (up to 30 meters), although the top of the Point Pleasant was still placed at the base of the Kope Formation. Thus it is clear that these authors included all of the Bromley Shale in their thickness tabulation.

Marker units in the Point Pleasant have been identified in detail and correlated widely in outcrop and core by the excellent work of McLaughlin & Brett (2001). Using a number of distinct marker horizons in the Point Pleasant and in the base of the overlying Fulton member of the Kope Foramtion, these authors have shown in detail the stratigraphic development of the northern Lexington Platform in northern Kentucky and southwestern Ohio. Key marker horizons included hardgrounds and faunal epiboles. Commonly hardgrounds in the Point Pleasant are developed at the top of storm dominated, meter-scale cycles. These hardgrounds often show evidence of early cementation of coarse grainstone cycle tops, numerous boring assemblages, and are often demarcated by phosphate and iron-stained pyrite mineralized surfaces. In addition to the endofauna, these hardgrounds are commonly colonized by epifaunal that produced attachment structures (i.e. *Anomalocrinus* and bryozoan holdfasts). In some cases, these hardgrounds were developed onto sculpted and channeled surfaces with evidence of overhanging ledges and development of numerous solutional cavities. At least one horizon shows evidence for erosional scouring and undercutting to form cryptic cavities that can be colonized by diverse faunas.

As for epiboles, immediately above the Point Pleasant are two remarkable horizons that contain very restricted occurrences of *Triarthus* and *Cheirocystis fultonensis.* The deep-water trilobite, *Triarthrus*, appears within two thin horizons in the basal Fulton in the area of Cincinnati and the unique rhombiferan cystoid, *Cheirocystis fultonensis*, has only been located within a single thin interval in the Fulton beds. The latter occurs just about a half meter above the top of the Point Pleasant. Additional markers that are somewhat discontinuous, but show some potential for further consideration include the K-bentonite horizons recognized by Schumacher and Carlton (1991) and referred to as the Bear Creek K-bentonites. It is likely that this succession of K-bentonites correlates with those in the upper Dolgeville to basal Indian Castle Shale of New York (after Baird & Brett, 2002). The hardground-rich interval at the top of the Point Pleasant, as in New York State at the top of the Steuben Limestone, represents the last phase of major widespread limestone-dominated deposition in the GACB. Thereafter, much of the GACB region became significantly influenced by flysch-style deposition associated with major continent-ward progradation of foreland infilling facies. Thus the Point Pleasant Limestone represents the last relatively pure limestone unit of the GACB.

Chapter 6: Time-Restricted Facies in the Upper Ordovician (Ashbyan-Mohawkian) of Eastern Laurentia and their Paleoenvironmental Significance

ABSTRACT

Time-restricted facies have been defined previously as distinctive stratigraphic units developed during specific temporal windows impacted by a range of possible environmental disturbances. As such, these facies are thought to record unique environmental perturbations that influenced physical, chemical, and biological processes that resulted in their deposition. Time-restricted facies have been documented for intervals in the Devonian (see Walliser 1984, 1986), but are described here for the first time for the Upper Ordovician rocks of eastern North America.

This study documents approximately six facies that appear to represent what might be referred to as environmentally sensitive time-restricted facies. These include: 1) interbedded, nodular chert to chert-impregnated, burrowed intervals, 2) condensed, greenish-gray, glauco-chamositic quartz-rich dolomite and shale intervals, 3) restricted marine to hypersaline carbonate facies with cross-bedded oolitic grainstones, stromatolitic boundstones, and evaporite crystal laths, 4) intervals of polymictic intra-and extra-clastic limestone beds, 5) laminated to rhythmically-bedded calcisiltite and shale facies that are prone to deformation and often contain numerous correlatable deformed horizons, and 6) a number of unique chemostratigraphic events. These facies appear to represent periodic changes in sea-level, ocean chemistry, temperature, salinity, aridity, runoff, and even pulses of tectonic episodes, etc. within the epeiric sea of eastern North America during this time interval.

The time-restricted facies documented herein support a “layer-cake renaissance” approach for use in foreland basin studies. Recognition of these facies provides for a

substantially more detailed record of Late Ordovician environmental change than reported previously. The linkages between widespread lithologic change, changes in sedimentary provenance, sea-water chemistry, sea-level change, and associated pulses in tectonic activity are more strongly linked. Hence, many of these events point to allocyclic processes that influenced deposition across much of the GACB, while a number of key tectonic events are also evident.

INTRODUCTION:

Many biological events are coordinated with distinct changes in lithology, sedimentology, and depositional facies (Walliser 1984, 1986). Moreover, many of these litho-events occur within a limited portion of a biozone and thus document a narrower time increment than the associated biotic event. Walliser (1986) and Lottman and colleagues (1986) suggest that these litho-events can often be recognized over wide regions, if not globally, within relevant facies and can be used for refined correlations. These authors supported the idea that some of these facies were produced under a range of conditions that were unique to a given time interval and thus produced only during those times. In other words certain facies do not exist continuously (in the strict Waltherian sense), but are time-restricted in their occurrence.

In the Late Ordovician (Ashbyan to Mohawkian) of Eastern Laurentia, there are a number of distinct, although sometimes subtle, facies that are emerging as potentially important events in the last stand of the Great American Carbonate Bank (GACB). These facies appear within relatively narrow time windows and can be developed in multiple outcrop regions of the GACB at roughly the same time. Such facies are often very similar (although not necessarily identical), and most are subtle, relatively thin, and occasionally discontinuous within local outcrop regions owing to associated unconformities. In some cases, these time-restricted facies have been recognized by previous authors as local marker horizons. However many of these facies have not been considered as regionally extensive and have been considered strictly parochial in construct. Nonetheless, although these facies can be relatively unremarkable locally, their occurrence elsewhere suggests the possibility of time-specific facies as suggested by Walliser (1996), especially when they can be constrained within relatively narrow time slices using other chronologies including event markers, biostratigraphy, and sequence stratigraphic approaches.

This study has focused on the Chazy, Black River and Trenton Groups and their equivalents across the eastern GACB where a number of facies and/or unique chemostratigraphic events appear to be developed within relatively narrow time intervals. These time-specific facies include: 1) interbedded, nodular chert to chert-impregnated, burrowed intervals, 2) condensed, greenish-gray, glauco-chamositic quartz-rich dolomite and shale intervals, 3) restricted marine to hypersaline carbonate facies with cross-bedded oolitic grainstones, stromatolitic boundstones, and evaporite crystal laths, 4) intervals of polymictic intra-and extra-clastic limestone beds, 5) laminated to rhythmically-bedded calcisiltite and shale facies that are prone to deformation and often contain numerous correlatable deformed horizons, and 6) a number of unique chemostratigraphic events.

Recognition of these facies in more than one outcrop region of the Upper Ordovician of eastern Laurentia suggests that these facies were developed as a direct result of allocyclic processes that influenced their deposition. Development of these time-specific facies further suggests important changes in climate, sea-level, and or paleoceanographic regimes that are as yet unrecognized for this critical time period. In addition, the relatively high-frequency of these time-specific facies, as suggested by Walliser (1984, 1986), may provide opportunities to decipher the relative impact of periodic, repeating events (i.e. climatic cycles on short and longer-term time scales) on evolutionary events. The following discussion focuses on these events and their potential implications.

GEOLOGIC BACKGROUND

The lower Upper Ordovician Chazy, Black River and Trenton Groups of northeastern North America consist of some 100 to 500 m of highly fossiliferous carbonates and interbedded shales (Keith, 1988). These rocks record a dramatic and rapid change in depositional setting from widespread carbonate platform environments of the Chazy-Black River into a deep foreland basin during deposition of the Trenton Group and Utica Shales. This change was coincident with a recognized period of global sea-level rise (Tippecanoe Supersequence of Sloss, 1963) and coincident with the onset of the Taconic Orogeny. In recent years there has been debate as to how much of the Tippecanoe transgression is attributed to actual sea-level rise versus subsidence owing to plate collision during the orogeny.

Associated sedimentologic changes in the GACB during this interval have generally been attributed to the subduction of the eastern edge of North America (**figure 1**). The collision

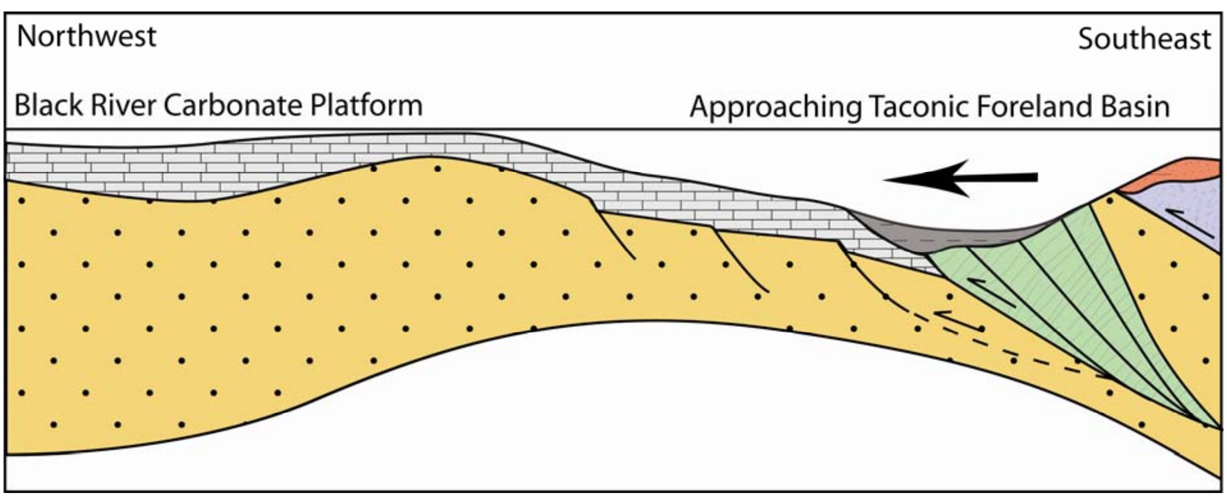


Figure 1: General model for the onset of the Taconic Orogeny and propagation of the Taconic Foreland Basin from southeast to northwest over a southeast-dipping subduction zone. Modified after Shanmugam and Lash, (1982)

initially generated a deep oceanic trough at the leading edge of the overriding plate (Iapetan

Ocean Plate). Eventually the trough migrated onto the Laurentian continental margin forming a peripheral style foreland basin (Fisher, 1962, 1977; Shanmugam & Lash, 1982; Lehmann et al., 1994). This effectively led to the demise of the GACB across wide regions of southeastern Laurentia. This collision also may have produced forebulges and back bulge basins distant from the foreland basin that may or may not have migrated laterally in time and space (Holland & Patzkowsky, 1997; Ettensohn, 2002b). The event may also have produced localized block movement in the vicinity of ancestral basement faults that reactivated in the collision thus impacting the topography of the platform and ultimately the architecture and function of the carbonate factory (Ettensohn, 1994).

The transition from the Black River into the Trenton has historically been the primary biostratigraphic and lithostratigraphic event of the CBRT interval. This boundary is associated with the widespread influx of shale from the Taconic Highlands and associated tectonic-induced subsidence during the main phase of the Taconic Orogeny. However, it has become clear that earlier events also had important roles in the development, distribution, and deposition of Chazy and Black River Group rocks. It has been known that the Taconic Orogeny was not a single event but a protracted period of tectonic activity along the Laurentian margin (Rodgers, 1971). Taconic tectonism has been defined as occurring in four tectophases – two of which are defined as major basin forming and filling episodes (**figure 2**). The earliest of these basin-forming and filling-phases, the Blountian Phase, activated first in the Ashbyan and influenced deposition through the Turinian. This tectonic episode appears to have been relatively small in its overall impact. In contrast, the second basin forming and filling episode, the Vermontian Phase activated in the mid-Mohawkian during deposition of the Trenton Group (Cisne et al., 1982) and impacted a significantly broader region. The pattern of sedimentologic and stratigraphic change

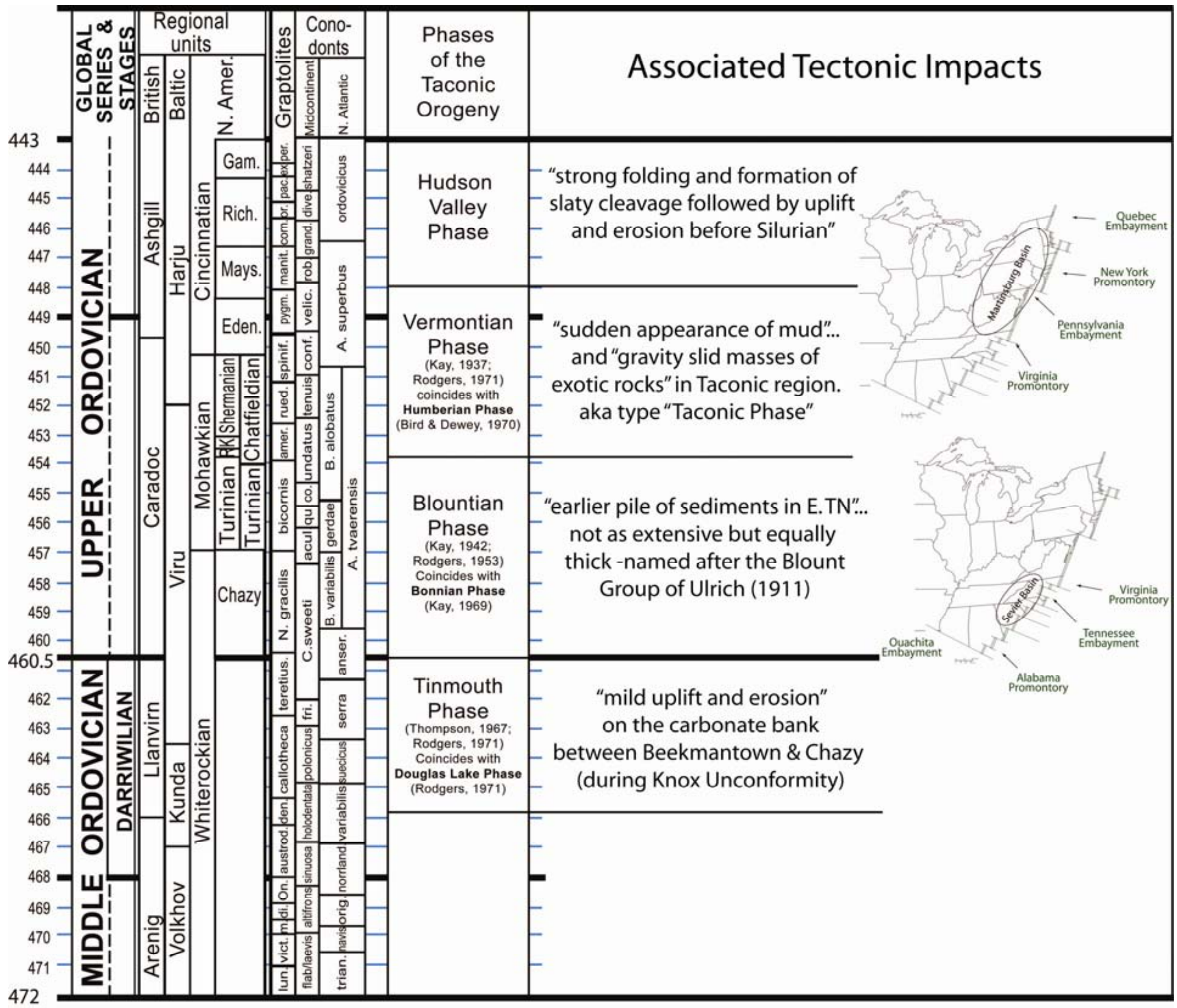


Figure 2: Middle and Upper Ordovician timescale for eastern North America (modified after Webby et. al, 2004). Included are the relative positions of previously identified phases of the protracted Taconic Orogeny. Four major phases have been recognized on sedimentologic and structural evidence including basin filling episodes which include both the Blountian and Vermontian phases.

in the epicratonic seas was quite likely to have been impacted by local and regional tectonism. What is not yet clear however is the degree to which these structural modifications impacted extrinsic controls on sedimentation including relative sea-level change, water mass circulation, climatic regimes, etc. and at what times these may have influenced deposition across the GACB.

STUDY REGION & GENERAL STRATIGRAPHIC REVIEW

The Chazy, Black River and Trenton groups (Vanuxem, 1838, 1842; Emmons, 1843) were among the first stratigraphic units to be described and named on a lithologic basis in North America. Collectively they make up the latest Champlainian and Mohawkian Stages of the Upper Ordovician. Subsequent to the surveys of New York, other survey geologists applied the New York-based nomenclature across much of eastern North America (**figure 3**) on the basis of

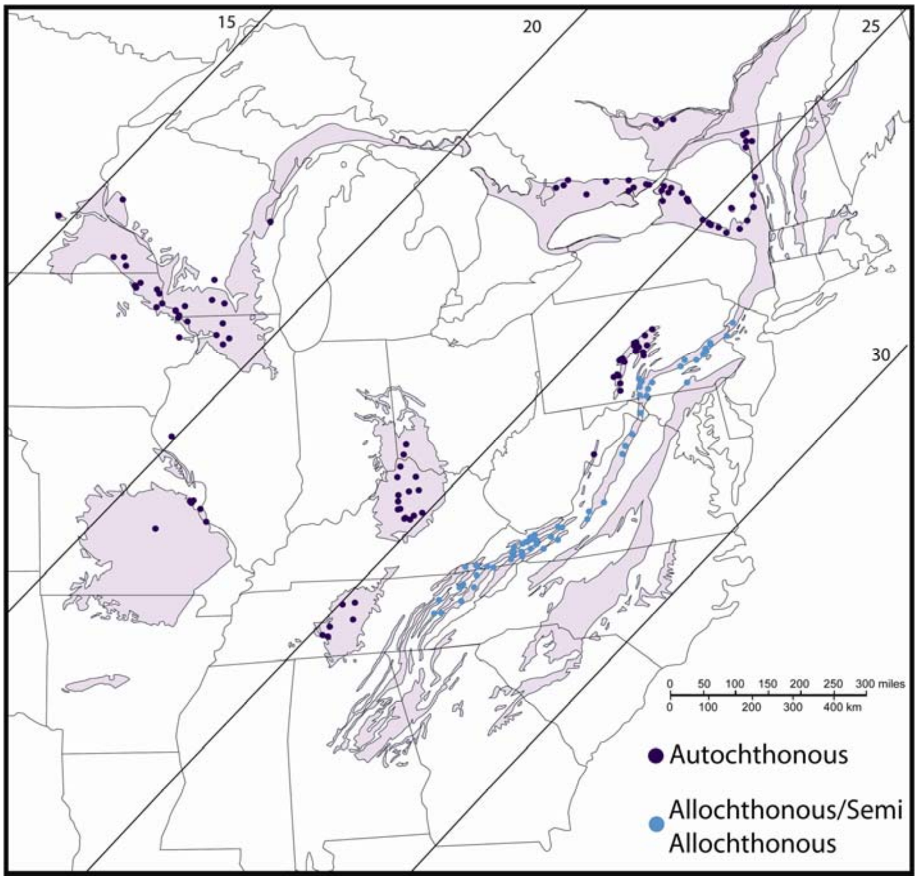
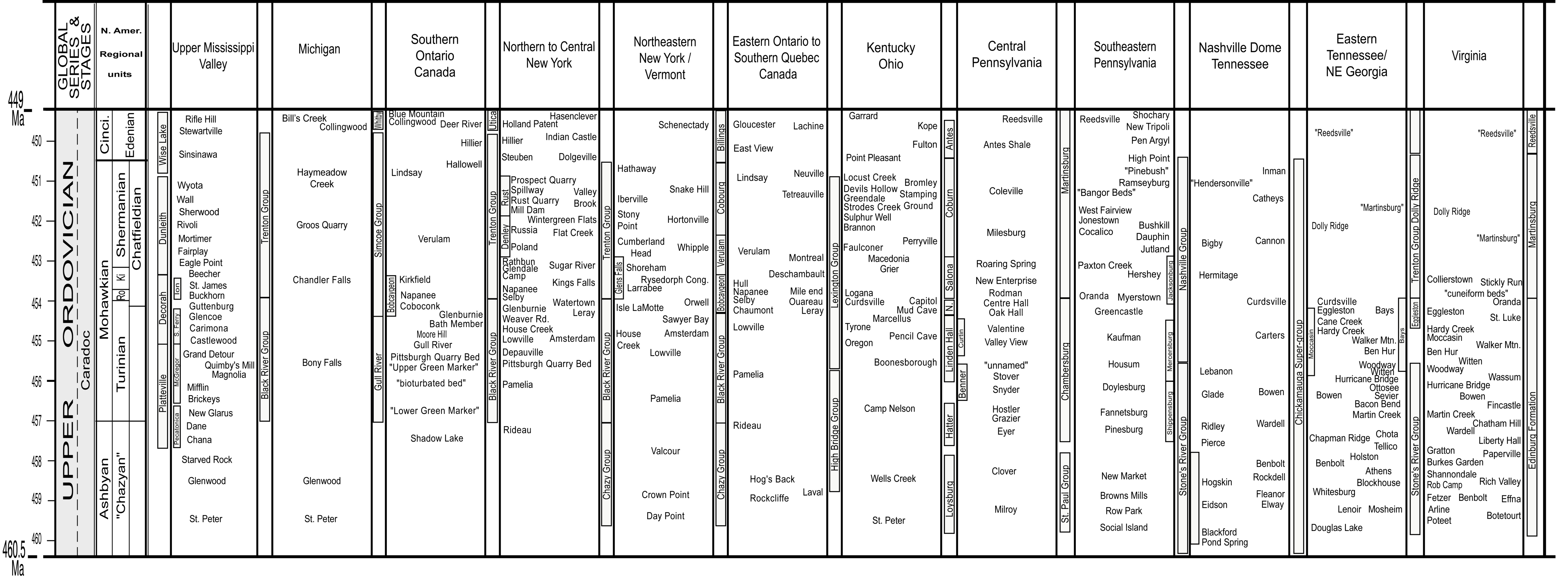


Figure 3: Ordovician outcrop distribution in eastern North America (shaded) with location of type localities for different Mohawkian aged strata (dots) and distinguished by their structural context (i.e. allochthonous – indicates outcrop localities that have been disturbed tectonically and have been folded, faulted and transported to some extent from their original depositional location). Relative paleolatitudinal position (in˚South) is also demarcated after Scotese and McKerrow (1990), and Witzke (1990).

gross lithology and fossil content. The collective application of biostratigraphy, lithostratigraphy and other principles helped these early workers develop a holistic view of these rock units as time parallel units. This concept is similar to what Walliser (1986) would consider as a "holostratigraphic" approach to stratigraphy. Subsequent work in disparate outcrop regions of the eastern U.S. and Canada led to further differentiation of these rocks with the result that each outcrop region became isolated and parochial in its nomenclature. Numerous local names (**figure 4)** have been applied to describe distinct biostratigraphic, lithostratigraphic, and chronostratigraphic intervals which, in most cases where used only on local scales or were not fully defined owing to incomplete exposure and/or differences in stratigraphic techniques. Moreover, the work in one region was often carried out irrespective of other outcrop areas. Thus numerous stratigraphic names were installed, most often producing significant synonymy and overlap with other regions. Thus, ultimately this historic practice has produced a highly complex nomenclature that continues to cloud true stratigraphic correlations (whether temporal or spatial) between seemingly disparate outcrop regions. Figure 4 illustrates the range of stratigraphic names, which for the purpose of correlation has presented substantial barriers for both historic and recent stratigraphic work. This problem has hindered basin analysis studies, and has exacerbated exploration and development of the highly productive Trenton-Black River natural gas plays. Recent publications have suggested the need for an improved stratigraphic nomenclature that can be applied across the Appalachian Basin region. Evenick and Hatcher (2006 a, b) have proposed that the original Black River –Trenton nomenclature take precedence across the region.

In order to make progress toward delineating the stratigraphic context of time-specific facies, the following discussion presents a basic overview of each outcrop region in the study area. The discussion establishes the basis for the stratigraphic correlations used in this study, and delineates a chronostratigraphy for use in assessment of geospatial and temporal changes in the development of rock units within the study region. Thus, this study compares outcrop sections from three primary regions: 1) the New York-Ontario type sections for the Upper Ordovician Chazy, Black River, and Trenton groups, 2) the Cincinnati Arch region where the High Bridge-

Figure 4: Stratigraphic nomenclature applied to rock units of the early Late Ordovician for twelve distinct regions of eastern North America. More than 300 stratigraphic names have been applied to rock units including and equivalent to the Chazy, Black River, and Trenton Groups.



GLOBAL SERIES & STAGES
N. Amer. Regional units
449 Ma
450
451
452
453
454
455
456
457
458
459
460
460.5 Ma
UPPER ORDOVICIAN
Caradoc
Cinci.
Edenian
Mohawkian
Shermanian
Chatfieldian
Ro
Ki
Turinian
Ashbyan
"Chazyan"
Upper Mississippi Valley
Wise Lake
Dunleith
Decorah
Ion
S. Ferry
McGregor
Platteville
Pecatonica
Rifle Hill
Stewartville
Sinsinawa
Wyota
Wall
Sherwood
Rivoli
Mortimer
Fairplay
Eagle Point
Beecher
St. James
Buckhorn
Guttenburg
Glencoe
Carimona
Castlewood
Grand Detour
Quimby's Mill
Magnolia
Mifflin
Brickeys
New Glarus
Dane
Chana
Starved Rock
Glenwood
St. Peter
Trenton Group
Black River Group
Michigan
Bill's Creek
Collingwood
Haymeadow Creek
Groos Quarry
Chandler Falls
Bony Falls
Glenwood
St. Peter
Whitby
Simcoe Group
Bobcaygeon
Gull River
Southern Ontario Canada
Blue Mountain
Collingwood
Deer River
Hillier
Hallowell
Lindsay
Verulam
Kirkfield
Napanee
Coboconk
Glenburnie
Bath Member
Moore Hill
Gull River
Pittsburgh Quarry Bed
"Upper Green Marker"
"bioturbated bed"
"Lower Green Marker"
Shadow Lake
Utica
Trenton Group
Black River Group
Northern to Central New York
Hasenclever
Holland Patent
Indian Castle
Hillier
Steuben
Dolgeville
Prospect Quarry
Spillway
Valley Brook
Rust
Rust Quarry
Mill Dam
Wintergreen Flats
Denley
Russia
Flat Creek
Poland
Rathbun
Glendale
Camp
Sugar River
Napanee
Selby
Kings Falls
Glenburnie
Watertown
Leray
Weaver Rd.
House Creek
Lowville
Amsterdam
Depauville
Pittsburgh Quarry Bed
Pamelia
Rideau
Trenton Group
Glens Falls
Black River Group
Chazy Group
Northeastern New York / Vermont
Schenectady
Hathaway
Snake Hill
Iberville
Stony Point
Hortonville
Cumberland Head
Whipple
Shoreham
Rysedorph Cong.
Larrabee
Orwell
Isle LaMotte
Sawyer Bay
House Creek
Amsterdam
Lowville
Pamelia
Valcour
Crown Point
Day Point
Billings
Cobourg
Verulam
Bobcaygeon
Black River Group
Chazy Group
Eastern Ontario to Southern Quebec Canada
Gloucester
Lachine
East View
Neuville
Lindsay
Tetreauville
Verulam
Montreal
Deschambault
Hull
Napanee
Selby
Chaumont
Mile end
Ouareau
Leray
Lowville
Pamelia
Rideau
Hog's Back
Laval
Rockcliffe
Lexington Group
High Bridge Group
Kentucky Ohio
Garrard
Kope
Fulton
Point Pleasant
Locust Creek
Devils Hollow
Bromley
Greendale
Stamping Ground
Strodes Creek
Sulphur Well
Brannon
Perryville
Faulconer
Macedonia
Grier
Logana
Curdsville
Capitol
Mud Cave
Marcellus
Tyrone
Pencil Cave
Oregon
Boonesborough
Camp Nelson
Wells Creek
St. Peter
Antes
Coburn
Salona
N.
Linden Hall
Curtin
Benner
Hatter
Loysburg
Central Pennsylvania
Reedsville
Antes Shale
Coleville
Milesburg
Roaring Spring
New Enterprise
Rodman
Centre Hall
Oak Hall
Valentine
Valley View
"unnamed"
Stover
Snyder
Hostler
Grazier
Eyer
Clover
Milroy
Martinsburg
Chambersburg
St. Paul Group
Southeastern Pennsylvania
Reedsville
Shochary
New Tripoli
Pen Argyl
High Point
"Pinebush"
Ramseyburg
"Bangor Beds"
West Fairview
Jonestown
Cocalico
Bushkill
Dauphin
Jutland
Paxton Creek
Hershey
Oranda
Myerstown
Greencastle
Kaufman
Housum
Doylesburg
Fannetsburg
Pinesburg
New Market
Browns Mills
Row Park
Social Island
Jacksonburg
Mercersburg
Shippensburg
Nashville Group
Stone's River Group
Nashville Dome Tennessee
"Hendersonville"
Inman
Catheys
Bigby
Cannon
Hermitage
Curdsville
Carters
Lebanon
Glade
Bowen
Ridley
Wardell
Pierce
Benbolt
Rockdell
Hogskin
Fleanor
Elway
Eidson
Blackford
Pond Spring
Chickamauga Super-group
Moccasin
Eastern Tennessee/ NE Georgia
"Reedsville"
"Martinsburg"
Dolly Ridge
Curdsville
Eggleston
Bays
Cane Creek
Hardy Creek
Walker Mtn.
Ben Hur
Woodway
Witten
Hurricane Bridge
Ottosee
Sevier
Bowen
Bacon Bend
Martin Creek
Chota
Chapman Ridge
Tellico
Holston
Benbolt
Athens
Blockhouse
Whitesburg
Lenoir
Mosheim
Douglas Lake
Trenton Group
Dolly Ridge
Eggleston
Bays
Stone's River Group
Virginia
"Reedsville"
Dolly Ridge
"Martinsburg"
Collierstown
Stickly Run
"cuneiform beds"
Oranda
Eggleston
St. Luke
Hardy Creek
Moccasin
Walker Mtn.
Ben Hur
Witten
Woodway
Wassum
Hurricane Bridge
Bowen
Fincastle
Martin Creek
Chatham Hill
Wardell
Liberty Hall
Gratton
Paperville
Burkes Garden
Shannondale
Rob Camp
Rich Valley
Fetzer
Benbolt
Effna
Arline
Poteet
Botetourt
Reedsville
Martinsburg
Edinburg Formation

Lexington-Cincinnati Group rocks were deposited, and 3) the intervening region of central Pennsylvania. In addition, where possible by recognition of correlated event horizons, correlations with additional regions are considered. Outcrop sections in the New York – Ontario region are located in the Grenville structural province. In contrast, the Cincinnati Arch region straddles the Precambrian Grenville suture with the older portions of the craton (**figure 5**). The

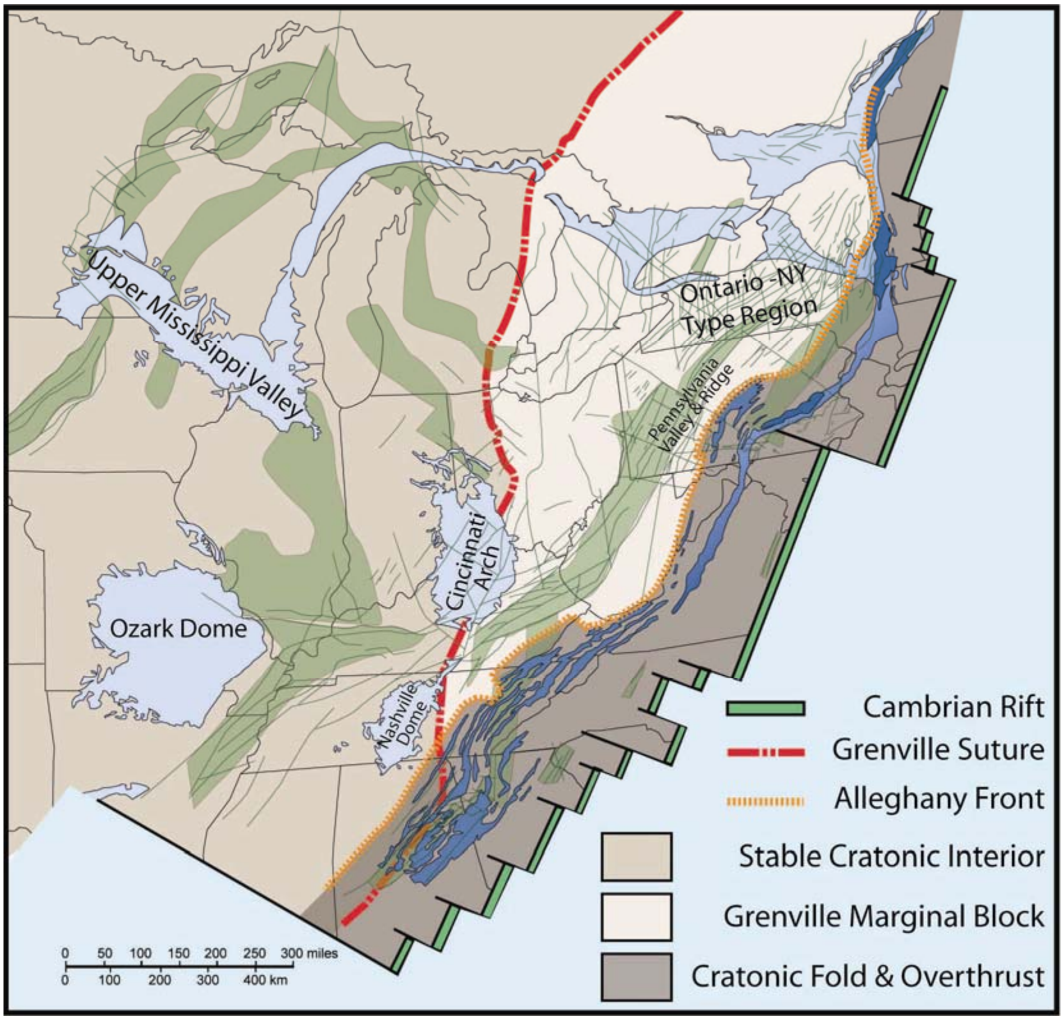


Figure 5: Cratonic setting for Ordovician outcrop regions in eastern North America. Within this region there are 3 distinct structural provinces. The first two from west to east, termed "Stable Cratonic Interior," and "Grenville Marginal Block" were emplaced during the Precambrian and had subsequently rifted along the Cambrian Rift margin during the latest Precambrian to early Cambrian. Within this region, numerous structural troughs are known to have filled with large thicknesses of siliciclastic sediment. The Great American Carbonate Bank was subsequently developed across the top of these structural provinces during the Cambro-Ordovician. Subsequently, to their deposition, the cratonic margin was over thrust beginning in the Ordovician Taconic Orogeny, and continuing intermittently through subsequent orogenies including the Silurian (Salinic), Devonian (Acadian), and Pennsylvanian (Alleghenian) that finally closed the ancestral Atlantic Ocean.

Cincinnati Arch region is located some 300 miles inboard of the cratonic margin and well inside the GACB. The NY-Ontario region runs roughly perpendicular to the structural margin of the Laurentian craton, and extends from near the margin itself to some 400 miles toward the interior of the GACB. These locations are themselves separated by some 500-600 miles. Outcrops in central Pennsylvania are afforded by the Ridge and Valley Province. Both of the former outcrop regions run perpendicular to slightly oblique to paleodepositional slopes.

For the Cincinnati Arch region, activation and subsidence of the intracratonic Sebree Trough is coincident with the Vermontian Phase of the Taconic Orogeny and produced a deepening slope to the north and west of the central Kentucky region. In New York and Ontario, relict topographic highs in the vicinity of the Adirondack and Frontenac Arch regions produced a yoked basin with a topographic high centered in central New York. Initially (during Chazy-Black River deposition) this topographic high yielded depositional slopes that deepened to the south and east toward the New York Promontory and the cratonic margin. A second depositional slope, albeit much less pronounced, deepened to the west. To the northeast of the Adirondack and Frontenac Arches was an additional basin, referred to commonly as the Ottawa Embayment.

As is the case in New York, the outcrop region of central and south-central Pennsylvania is characterized also by what appears to be a yoked basin. Deposition during the Chazy-Black River interval (Ashbyan-Turinian) appears to have been influenced by a topographic high referred to as a southern extension of the Adirondack Arch (see Ch. 1, Fig. 22). This feature was located some distance inboard from the cratonic margin. It too produced a northwesterly facing depositional slope, dipping toward the craton interior. Likewise, as is the case in New York, there is also a southeasterly dipping depositional slope – again toward the cratonic margin. This feature appears to have been a pronounced topographic high at different times during the Ashbyan-Turinian and it appears to have been finally breeched during the deposition of the

Nealmont through Salona interval.

Chronostratigraphic Framework for New York and Ontario:

Recent studies in the type region have established the ability to use: among other things, physical stratigraphy, graptolite, and conodont biostratigraphy, event beds including K-bentonites, taphofacies correlations, and sequence stratigraphic models to develop refined chronostratigraphic correlations for the Black River and Trenton groups in their type region (Goldman et al, 1994; Mitchell et al, 1994; Baird et al, 1992; Brett and Baird, 2002; Baird and Brett, 2002; Armstrong et al, 1994; Melchin et al, 1994; Brookfield; Cornell, 2001; Brett et al., 2004) (**figure 6**). Through these studies, it has become apparent that a variety of unique horizons and distinctive lithologic intervals are recognizable over wide portions of the New York-Ontario outcrop belt and provide a basis for correlation both into the subsurface and into other outcrop regions.

Chronostratigraphic Framework for the Cincinnati Arch Region:

Another region that has received recent study is the Jessamine Dome to Nashville Dome region in Kentucky and Tennessee (**figure 7)**. Detailed litho-, bio-, event and sequence stratigraphy has been established in the High Bridge Group and Lexington Limestone and the Carters and Hermitage groups of Tennessee (Cressman, 1973; Cressman & Noger, 1976; Kolata et al., 1996, Holland & Patzkowsky 1996, 1998; Kolata et al., 1996; Brett et al, 2004). In this region, these strata are divisible into several third order sequences (M1 to M6 of Holland & Patzkowsky, 1996). Progress has also been made toward developing a high-resolution cyclostratigraphy for the upper Mohawkian and Cincinnatian strata from central Kentucky through western Virginia (Pope & Read, 1995), and although there is some disagreement to cycle boundaries and

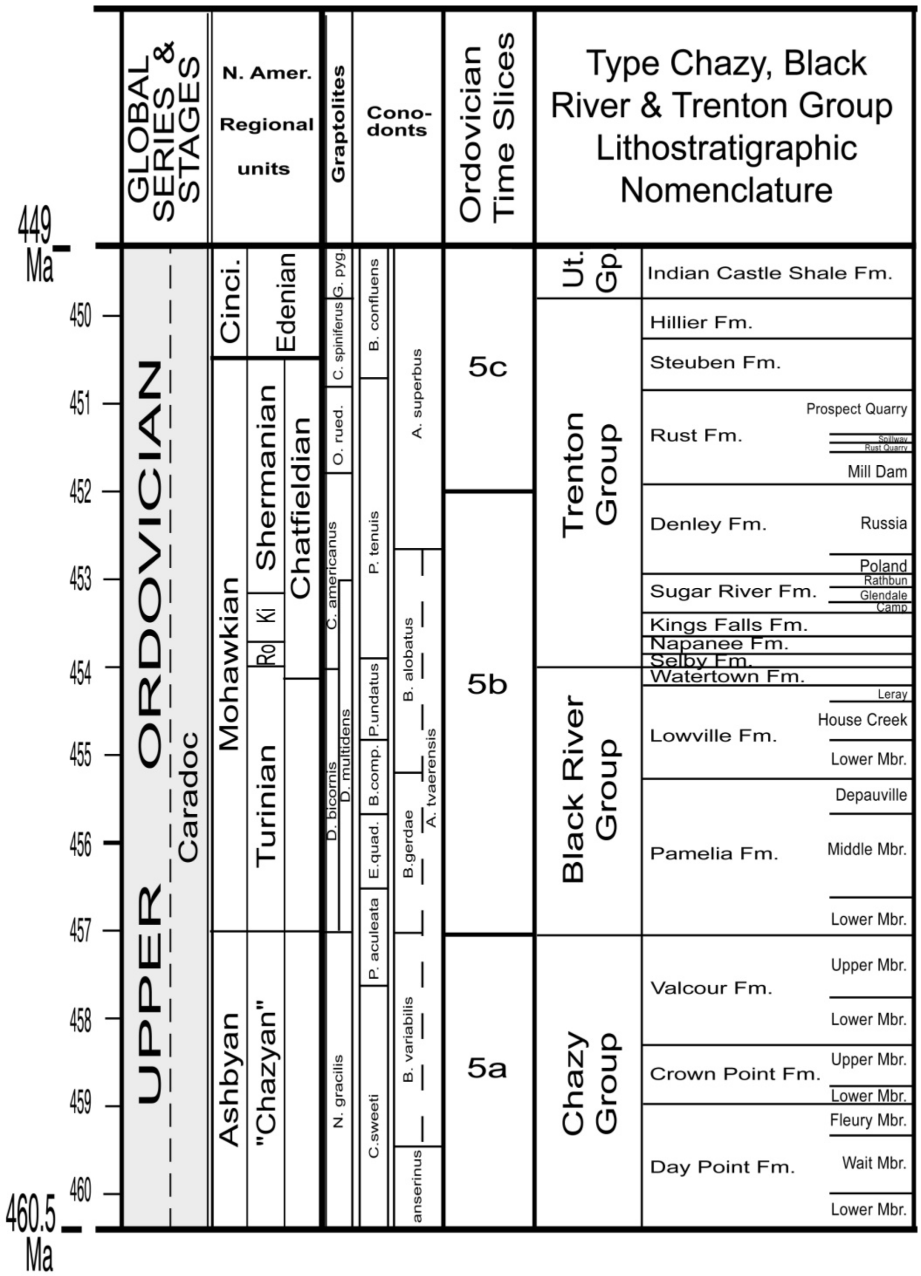


Figure 6: Relative and absolute Time Scale for the Upper Ordovician rocks of New York State as assigned in this study. Included on this diagram of Upper Ordovician rocks are the major lithostratigraphic intervals from the type region of the Mohawkian Stage as well as the key biostratigraphic zones and Ordovician time slices as recognized by Webby et. al, (2004) and applied to New York by the current author.

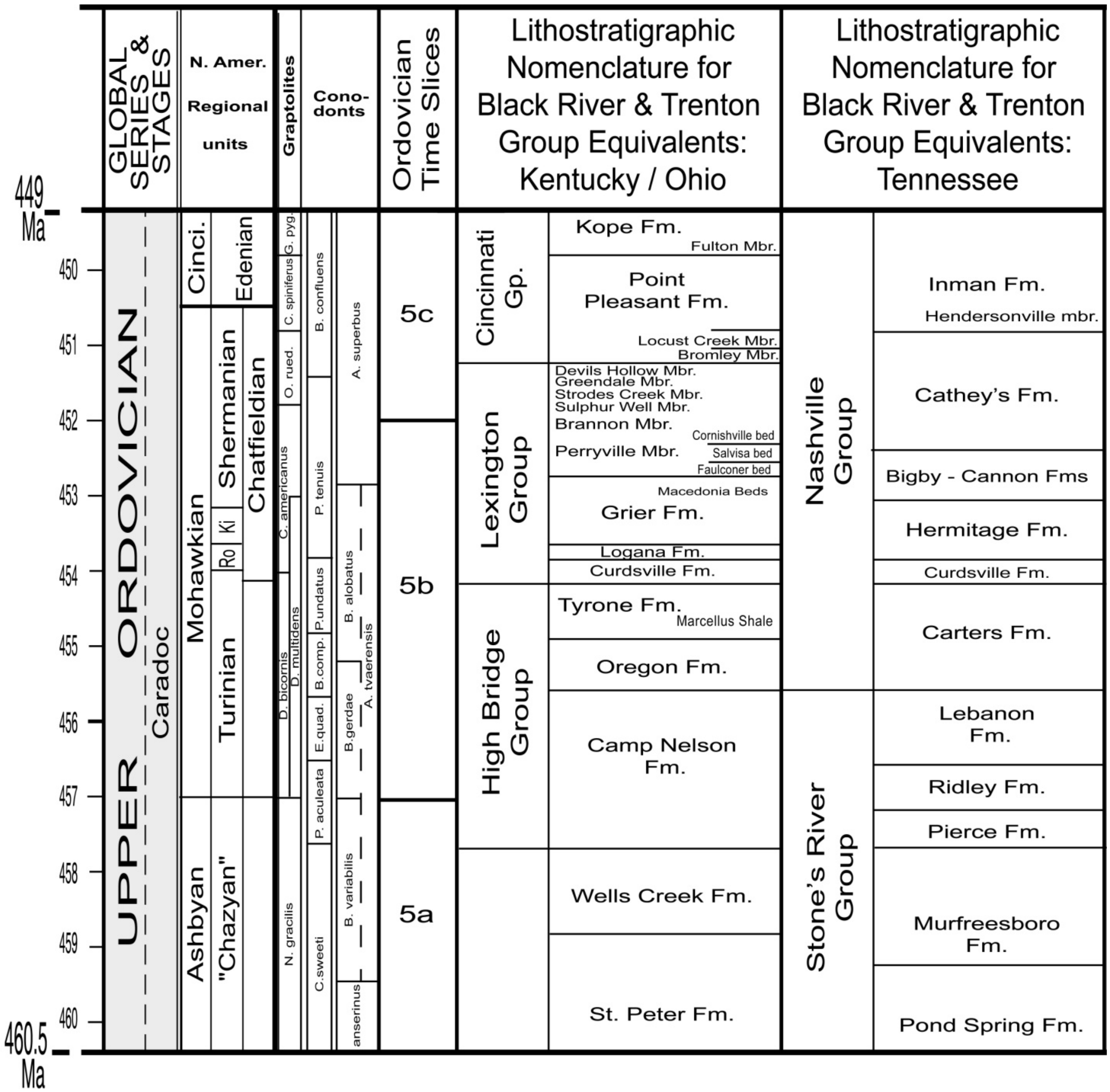


Figure 7: Relative and absolute Time Scale for the Upper Ordovician rocks of the Jessamine and Nashville Dome regions as assigned in this study. Included on this diagram of Upper Ordovician rocks are the major lithostratigraphic intervals from each region as well as biostratigraphic zones and Ordovician time slices as recognized by Webby et. al, (2004) and applied to herein by the current author.

composition (McLaughlin et al, 2001) lithologic packages are correlative across broad regions.

In addition to sequence and cyclostratigraphic studies, a number of event-stratigraphic studies also help to identify and establish local to regional tie lines. For instance, work by Haynes (1994) and Kolata and colleagues (1996) demonstrated the unique chemistries of the

Deicke and Millbrig K-Bentonites on the Cincinnati Arch and in the Valley and Ridge Province, including West Virginia and Virginia. In addition to K-bentonites, research on C-isotope stratigraphy in this region (Saltzman et al., 2001; Young et al., 2005) provides the basis for additional stratigraphic correlations and thus helps establish a baseline for this study.

Chronostratigraphic Framework for southern and central Pennsylvania:

Another region critical in the understanding of Upper Ordovician rocks of eastern North America is the Central Pennsylvania region. **Figure 8** shows the lithostratigraphic nomenclature as used in this study. Outcrop regions are divided into north-western central Pennsylvania and those rocks exposed in the Great Valley region roughly from the West Virginia - Maryland border to the Hershey, Pennsylvania region. This lithostratigraphy is based on lithostratigraphic studies of Kay, 1944, Thompson, 1963, Wagner, 1966, Rones, 1969, Berg, 1980, and studies herein. Biostratigraphic framework here is consistent with studies by Sweet & Bergstrom (1976), Sweet (1988), and Ryder and colleagues (1992). To date, no sequence stratigraphic studies have been published on the outcrop region as a whole, although some initial work has been presented by Slupik (1999), and Laughrey and colleagues (2004) for a limited number of outcrops in the western fold and thrust belt region. Additional contributions to distinguishing the Black River – Trenton Group boundary interval have been made by Barta (2004, 2006).

In terms of regional correlations, Wagner (1966) extended correlations, including the excellent work of Thompson (1963), from outcrop into the subsurface. Although units are coarsely defined, Wagner demonstrated the continuity of critical stratal units along two main transects. The first runs roughly NW/SE along the Alleghany Plateau and the second along a parallel transect through the Appalachian fold and thrust belt about 200 km to the southeast of

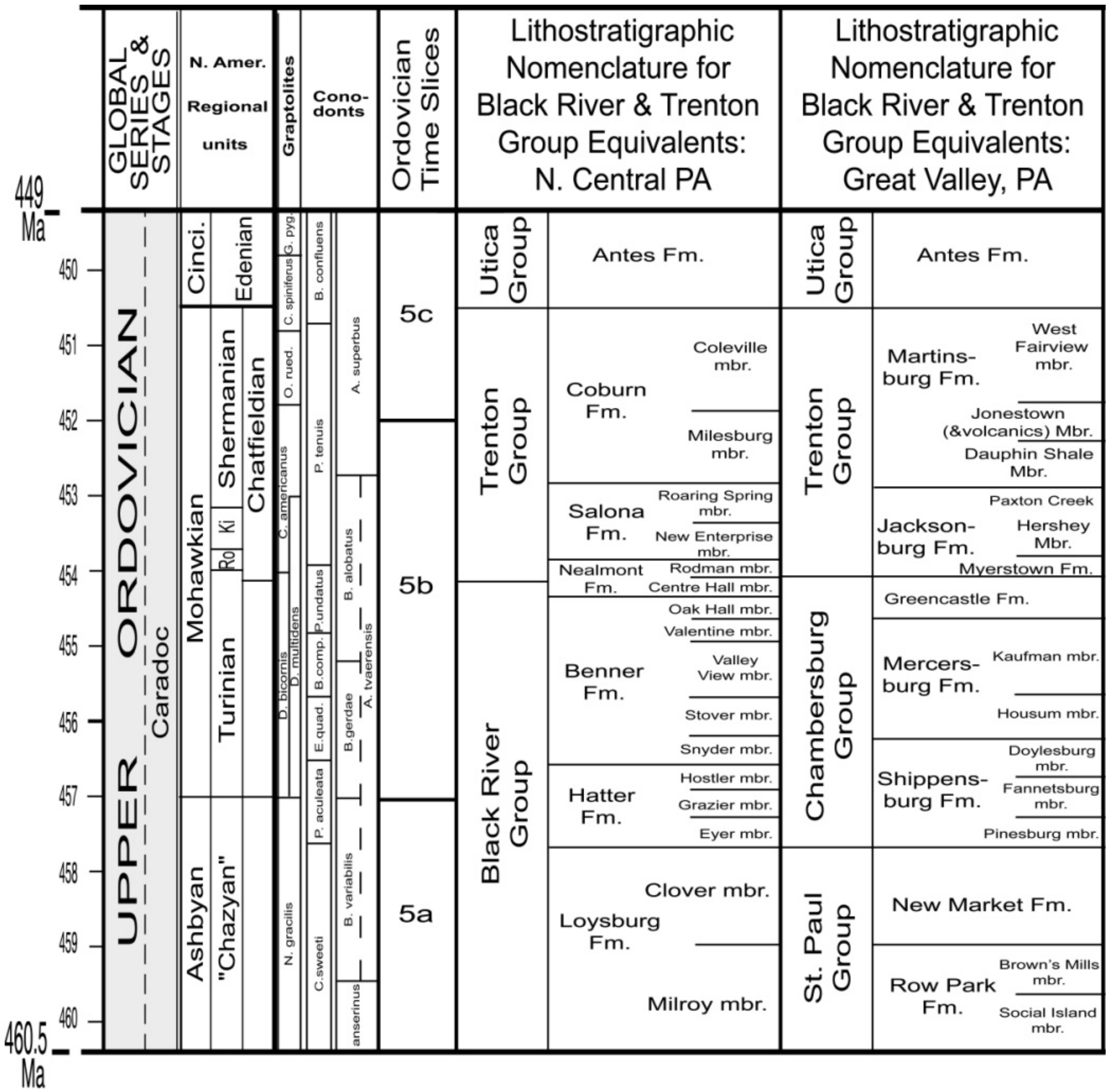


Figure 8: Relative and absolute Time Scale for the Upper Ordovician rocks of the Central Pennsylvania region as assigned in this study. Included on this diagram of Upper Ordovician rocks are the major lithostratigraphic intervals from the westernmost Appalachian Fold and Thrust belt, and the easternmost region of the fold and thrust called the Great Valley.

the first. Moreover, Wagner (1966) also extended his correlations into both southwestern New York/Ontario and into eastern Ohio/West Virginia.

TIME-RESTRICTED FACIES OF THE CBRT

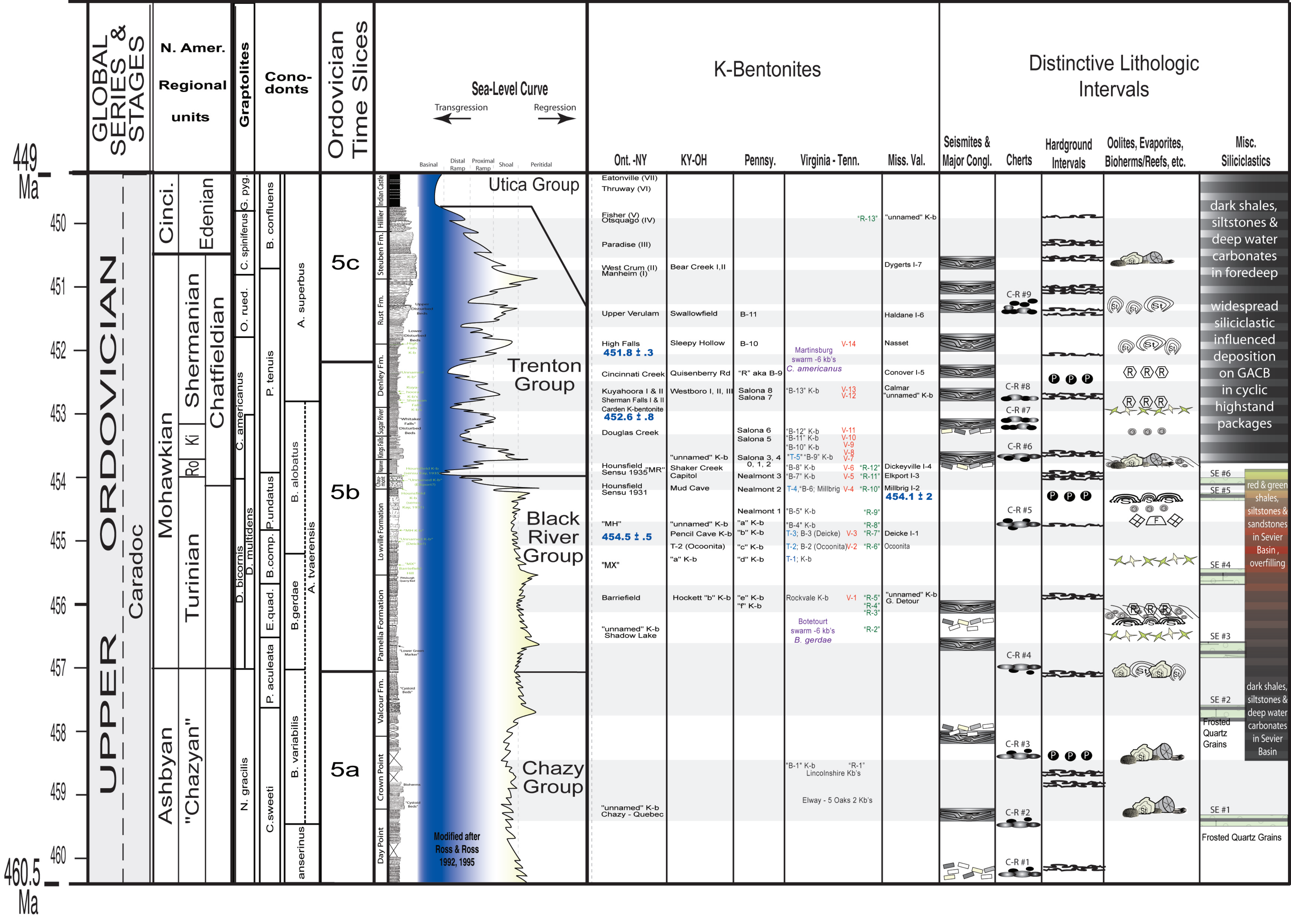
During the detailed re-evaluation of numerous historic stratigraphic studies herein, a number of important traditionally-defined event markers as well as unique lithologic markers emerged as possible candidates for lateral time-parallel correlations. Although these units are usually not the dominant lithologic units found within the CBRT interval, their subtle reappearance at seemingly predictable horizons relative to other event markers has required a detailed evaluation of their stratigraphic context (**figure 9**). Termed here as distinctive lithologic intervals, these appear to be analogous to time-restricted facies as suggested by Walliser (1984, 1986). The following section describes the dominant lithofacies and their occurrences within the primary study regions and their lateral equivalents.

<u>Chert Production & Time Restricted Facies</u>

The first and one of the most obvious accessory lithologies within the CBRT interval is a diverse array of limestone-hosted cherts. In most cases, cherts within CBRT strata take the form of dark grey to black nodular to bedded chert horizons, or chert filled burrows. Cherts may also be white to light chalky grey and associated with vuggy dolomitic intervals or exceptionally pure fenestral micrite deposits (**figure 10**). The formation of chert is thought to result from a number of different processes including: biological production of silica (nannobacteria, diatoms, radiolarians, sponges, etc.), upwelling of deep marine waters with dissolved silica, volcanic ash demineralization, post-burial hydrothermal precipitation from dissolution of quartz-rich deposits (sandstones, siltstones, and some clays), hot spring precipitation, as well as recrystallization of wind-blown silica dust in aqueous environments (Folk, 2002; Tomescu & Kidder, 2002a, b; Gammon & James, 2002).

In most cases in the CBRT the processes of chert formation are still not well understood. Nonetheless, most previous studies in the Upper Ordovician carbonates have attributed the

Figure 9: Biostratigraphic and event stratigraphic framework (K-bentonites) for the Upper Ordovician of eastern Laurentia projected into the lithostratigraphic framework of the type Chazy, Black River, Trenton group composite stratigraphy. Also shown are the relative positions of distinctive lithologic intervals or "time-specific facies" of Walliser (1984, 1986).



GLOBAL SERIES & STAGES
N. Amer. Regional units
Graptolites
Conodonts
Ordovician Time Slices
Sea-Level Curve
Transgression
Regression
Basinal
Distal Ramp
Proximal Ramp
Shoal
Peritidal
K-Bentonites
Ont. -NY
KY-OH
Pennsy.
Virginia - Tenn.
Miss. Val.
Distinctive Lithologic Intervals
Seismites & Major Congl.
Cherts
Hardground Intervals
Oolites, Evaporites, Bioherms/Reefs, etc.
Misc. Siliciclastics
449 Ma
460.5 Ma
UPPER ORDOVICIAN
Caradoc
Cinci.
Edenian
Mohawkian
Shermanian
Chatfieldian
Turinian
Ashbyan
"Chazyan"
5c
5b
5a
Utica Group
Trenton Group
Black River Group
Chazy Group
Modified after Ross & Ross 1992, 1995
Eatonville (VII)
Thruway (VI)
Fisher (V)
Otsquago (IV)
Paradise (III)
West Crum (II)
Manheim (I)
Upper Verulam
High Falls
451.8 ± .3
Cincinnati Creek
Kuyahoora I & II
Sherman Falls I & II
Carden K-bentonite
452.6 ± .8
Douglas Creek
Hounsfield Sensu 1935
"MR"
Hounsfield Sensu 1931
"MH"
454.5 ± .5
"MX"
Barriefield
"unnamed" K-b Shadow Lake
"unnamed" K-b Chazy - Quebec
Bear Creek I,II
Swallowfield
Sleepy Hollow
Quisenberry Rd
Westboro I, II, III
"unnamed" K-b
Shaker Creek
Capitol
Mud Cave
Pencil Cave K-b
T-2 (Ocoonita)
"a" K-b
Hockett "b" K-b
B-11
B-10
"R" aka B-9
Salona 8
Salona 7
Salona 6
Salona 5
Salona 3, 4
0, 1, 2
Nealmont 3
Nealmont 2
Nealmont 1
Martinsburg swarm -6 kb's
C. americanus
Botetourt swarm -6 kb's
B. gerdae
Rockvale K-b
Lincolnshire Kb's
Elway - 5 Oaks 2 Kb's
"unnamed" K-b
Dygerts I-7
Haldane I-6
Nasset
Conover I-5
Calmar "unnamed" K-b
Dickeyville I-4
Elkport I-3
Millbrig I-2
454.1 ± 2
Deicke I-1
Ocoonita
"unnamed" K-b
G. Detour
C-R #9
C-R #8
C-R #7
C-R #6
C-R #5
C-R #4
C-R #3
C-R #2
C-R #1
dark shales, siltstones & deep water carbonates in foredeep
widespread siliciclastic influenced deposition on GACB in cyclic highstand packages
red & green shales, siltstones & sandstones in Sevier Basin, overfilling
dark shales, siltstones & deep water carbonates in Sevier Basin
SE #6
SE #5
SE #4
SE #3
SE #2
SE #1
Frosted Quartz Grains

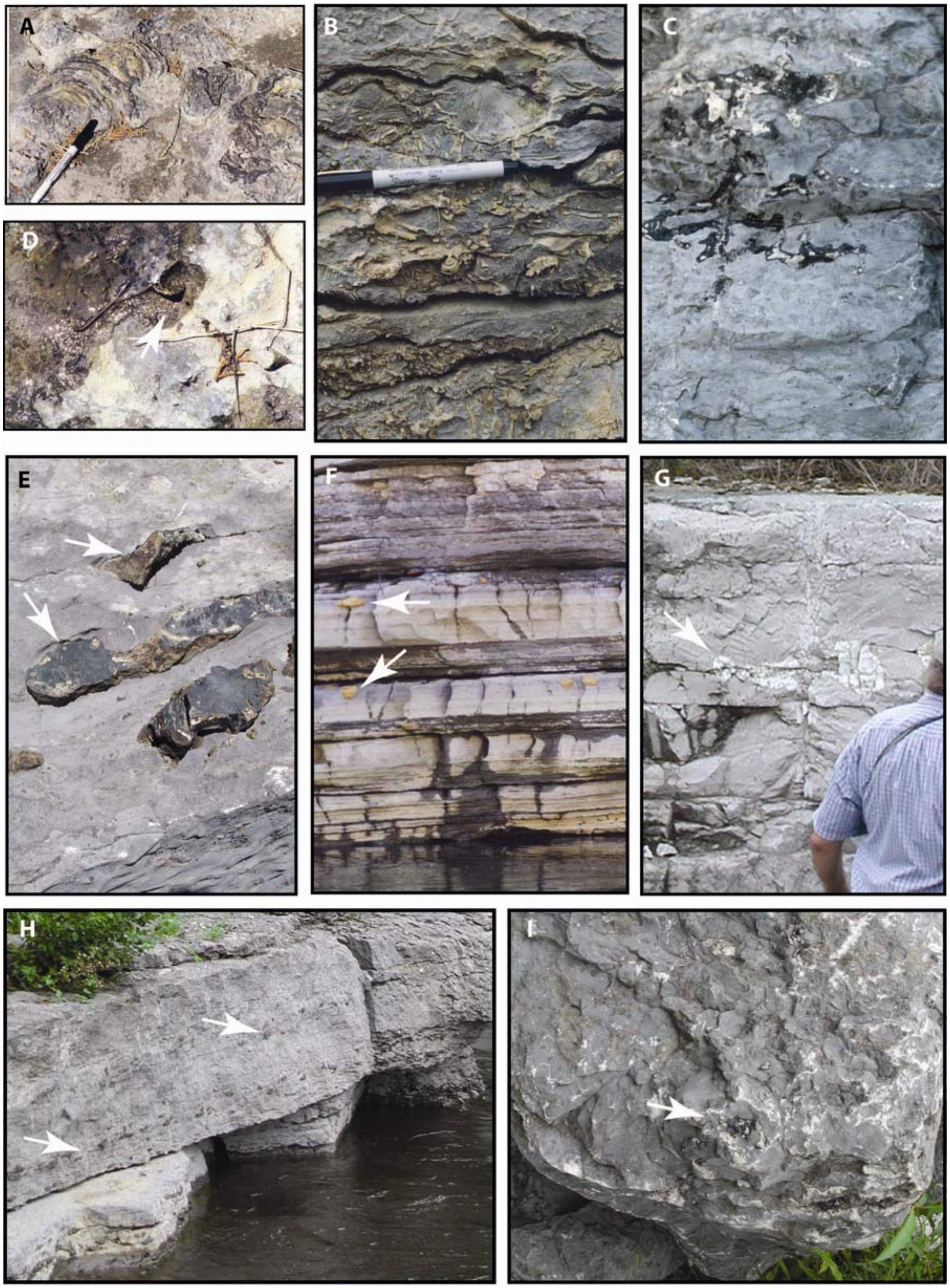


Figure 10: Examples of silicified fossil horizons and chert-rich intervals from a variety of intervals in the Chazy and Black River Groups. Included in the top row are a variety of silicified fossil assemblages including: A) silicified Stromatoporoids from the Watertown Formation, B) silicified Tetradium biostromal unit with silicified crinoid columnals, gastropods, brachiopods, bivalves, and vertical burrow infillings from the Lower Bobcaygeon of Marmora, Ontario, C) Chert-filled burrows, D) Silicified rugose corals *Lambeophyllum profundum* and *Streptelasma corniculum* both represented in the Watertown Formation. In

the second row are representative of some of the chert-rich facies found. E) Dark grey to black lenticular chert nodules (weathering in relief) from the Watertown Formation – these can become fairly uniform and bedded in some intervals in this unit such as are shown in image H, F) Buff-orange weathering white-chalky cherts from the House Creek Member of the Lowville Formation, G) Whitish-grey chert in a micrite matrix from the Tyrone Formation, High Bridge Group, Kentucky; and I) disseminated nodules of dark black chert in extremely bioturbated medium-grained wackestones of the Day Point Formation of the Chazy Group.

formation of cherts to a mixture of both primary deposition of silica (some cherts are composed of significant proportions of siliceous spicules and or early radiolarian tests), and early to late diagenesis (Steffen & Pope, 2002). The former authors suggest, on both chemical and physical criteria, that many cherts in the Upper Ordovician may be biogenic in origin (sponge-radiolarian sourced). Most are modified by early diagenesis, but maintain a uniform distribution characteristic of original depositional layering. A few cherts in CBRT strata have been attributed to late diagenesis where various minerals including evaporite minerals have been dissolved and replaced by cryptocrystalline silica (i.e. perhaps some lower-middle Dunleith cherts of the Iowa-Illinois Basin region; Witzke, 1987). Still other cherts appear to be associated with devitrification and alteration of volcanic ash or quartz-rich sandstones in marine environments and are commonly clustered above and below altered ash beds (Krekeler et al, 1994, Kolata et al., 1996; Marker & Huff, 2005).

Biotic Control on Silica/Chert Production?

Kidder and Erwin (2001) have noted the strong correlation through the Phanerozoic between radiations/extinctions of marine taxa (including the rapid diversification of sponges and radiolarians) and the number of chert-rich intervals. For the Upper Ordovician, there is an extremely rapid increase in the number of chert-rich intervals (3-4 fold increase over lower Ordovician) coincident with the rapid diversification of radiolarians and fairly modest diversification of glass sponges. In fact Kidder and Erwin suggest that radiolarians may have become the dominant biological control on the silica-cycle during this time. Following the end Ordovician mass extinction, the number of chert horizons drops very substantially for both silicified fossil occurrences and bedded-cherts – suggesting that, again, even in cases where cherts are clearly diagenetic – there might be a biological control on the original source of silica.

Regardless, it is now thought that filamentous nannobacteria may have a significant role in the precipitation of cherts such that even cherts that were previously believed to be abiotic or post-depositional, might be in some way biotically mediated during early diagenesis (Folk, 2002; Tobin, 2004) and therefore reflect some primary environmental control. Interestingly, the widespread distribution of chert-rich intervals in the CBRT interval appear to support a series of repeating events of limited time duration that modified the supply and availability of silica on the otherwise silica-starved carbonate platform of the GACB. In most cases their occurrence is tied to a combination of extrabasinal phenomena including perhaps climate, global sea-level change and associated oceanographic events including slight modification of water chemistry (slightly more acidic to slightly more alkaline), and/or volcanic/tectonic events that may have supplied important silica.

Where cherts are clearly of biogenic origin, the supply of dissolved silica is the main limiting factor controlling the distribution and abundance of silica secreting organisms. Quite commonly carbonate platforms and warm shallow marine settings away from siliciclastic source areas, and likely those of the GACB, have slightly elevated pH's and are silica-limited. Thus these environments tend to have a limited abundance of silica-secreting organisms relative to carbonate producers. Therefore, silica-producers rely on the influx of silica from an external source if the biota is to proliferate in the warm-shallow oligotrophic settings of a carbonate platform. This silica can be sourced from: 1) riverine input of dissolved silica from weathered siliciclastic rocks, 2) the alteration of particulate siliciclastic dust (including volcanic ash) in aqueous environments (depending on pH), and 3) upwelling of water from deep marine settings below the photic zone where dissolved silica has higher concentrations.

Upwelling Controls on Silica Delivery and Chert Production?

Upwelling of dissolved silica from the deep ocean is thought to be the main source for GACB biogenic cherts – especially along the southern Laurentian cratonic margin (Pope, 2002; Pope & Steffan, 2003). If upwelling is to bring significant quantities of dissolved silica (and other nutrients) from the deep ocean onto the craton and into the Laurentian epicontinental sea a mechanism is needed. Typically upwelling zones (and shallow mixing zones) form on continental shelves in stormy, low pressure belts or in regions where wind patterns are directed offshore. In these cases, surface waters are blown away from the continent which allows somewhat deeper waters to move toward the surface to replace the wind-driven water. Although this results in upwelling, the effect is typically seasonal and not sustained for significant periods of time, and very commonly the upwelling is produced offshore and over limited geographic areas. If there were seasonal upwelling in the Ordovician on the GACB, it would not likely have the capacity to impact large areas of Laurentia nor is it likely to have supplied a significant volume of dissolved silica for prolonged periods of time.

In other cases, sustained upwelling, even in equatorial areas inboard of continental margins, can occur if local submarine topography modifies oceanic circulation patterns. In these cases, vigorous thermohaline circulation is controlled by the architecture of the basin. Thus upwelling is induced where continental geometries restrict or otherwise force water masses to alter their flow direction. For the Ordovician, Kolata and colleagues (2001), based on paleogeographic reconstructions, postulated the impact of the formation of the Sebree Trough as an intracratonic conduit for such upwelling to occur. They suggest (along with Young et al., 2005) that a combination of factors may have resulted in establishment of upwelling in the GACB. Specifically, they advocate that Taconic tectonism reactivated the Reelfoot Rift and Rough Creek Grabens (**Ch. 1 figure 13**). This resulted in renewed subsidence and formation of the Sebree Trough. The orientation of the Laurentian craton and the "Trenton transgression" positioned the trough so that it was a negative feature that depressed below the thermocline. This geographic arrangement enabled the introduction of cool, nutrient-rich waters (phosphates, nitrates, and dissolved silica) into the GACB region beginning during the deposition of the Lexington Formation and continuing at least through the deposition of the Cincinnati Group when the GACB was effectively terminated.

Although this model has significant merit and may account for the source of silica for some of the chert occurrences noted herein, the timing and distribution of other chert beds is still problematic as some predate the development of the Sebree Trough, and in some cases cherts are either lacking or not as well developed along the Sebree as elsewhere. Moreover in most cases, the cherts are not associated with the deepest water facies where they might be expected. For instance earlier cherts are found associated with peritidal to very shallow subtidal limestones of the Black River Group and its equivalents. These were clearly deposited under substantially shallower and warmer water conditions as indicated by large photozoans (including dasycladacean algae), as well as a number of other indicators (Holland & Patzkowsky, 1996, 1998). The former authors have also suggested a linkage between sea-level lowstand, and chert development – whereby meteoric diagenesis during lowstands may help to facilitate the dissolution of silica-rich phases and reprecipitation of silica as cherts at or below sequence bounding unconformities provided there is substantial silica available for dissolution in the first place. This scenario appears to be plausible for some of these shallow-water chert occurrences, although not all cherts are located below an obvious sequence boundary - some are located in transgressive facies immediately above them.

Riverine Controls on Silica Delivery and Chert Production?

Thus if a majority of GACB cherts in correlatable intervals cannot confidently be attributed to upwelling in the Sebree Trough, an additional source of silica is needed. As mentioned, dissolved silica can be delivered through river influx in the vicinity of a carbonate platform provided the river is supplied with a significant source of groundwater relative to runoff (e.g. northeastern Australia). This is manifested in the requirement that particulate siliciclastic sediment loads need to be very low (high sedimentation rates are detrimental to the carbonate factory), but most importantly the riverine input needs to contain a significant supply of dissolved silica. In most humid tropical environments, even with higher rates of chemical weathering, dissolved silica concentration is low in rivers with high runoff-sourced discharge. However, in intermittently arid environments where runoff volume is low – relative to groundwater discharge, riverine discharge is either more acidic than fresh water (pH <6) or significantly alkaline (pH >8) and therefore can contain a significantly higher dissolved silica load. This silica then can flow out into marine settings where it will mix with marine waters and either precipitate as amorphous silica or be used by silica producing organisms. In these areas, typically deep water benthic faunas including hexactinellid sponges are known to become significantly more abundant in shallower water environments (Gammon, 2002).

In the case of the Ordovician GACB, given the reconstructed position of eastern Laurentia in the subtropical arid belt just south of the paleoequator, there is a potential for arid conditions to have developed over much of the exposed continent. Neodymium isotope data published by Fanton & Holmden (2007), suggest that a number of sea-level rise and fall events are recorded in the upper Mississippi Valley Dunleith Group strata. These are recognized by cyclic variation in the $^{\varepsilon}$Nd values. In this analysis, the introduction of Grenville Province or Taconic Mountain derived sediments to the Illinois Basin region raised $^{\varepsilon}$Nd values which could only occur during sea-level highstands. Conversely, the lowering of $^{\varepsilon}$Nd values suggests sediment contribution only from the local Superior Province sediments which would be transported into the Illinois Basin during sea-level lowstands. In the upper Mississippi Valley, and elsewhere, the occurrence of some chert horizons in shallow water (albeit often transgressive) facies is coincident with substantially more negative $^{\varepsilon}$Nd values (circa -20). This suggests that upwelling may not be the source of silica. Given the proximity of these regions to the Transcontinental Arch, together with the increased exposure of the arch during lowstands, it is possible that biogenic silica producers obtained silica – not from upwelling sources, but from riverine sources. The occurrence of these nodular cherts in transgressive intervals may be enhanced by rejuvenation of the carbonate factory, increased incidence of bioturbation, and the impact of storm processes. Collectively, these may also have acted to rework sediments and remobilize dissolved silica from pore waters (along with other nutrients) making them available again to silica sediment producers.

Diagenetic Mechanisms for Chert Production?

The final group of cherts documented in the GACB, are those that are thought to be formed during late diagenesis. This type of chert, both whitish-gray and dark grey to black, is commonly found in close proximity to volcanic ash beds. As reported by Krekeler and colleagues (1994) and Marker and Huff (2005), these chert beds may be superjacent and subjacent to relatively thick K-bentonite deposits. These cherts may cut across depositional layering, and commonly impregnate or silicify various fossils including corals and stromatoporoids. Within the CBRT, the biggest occurrence of K-bentonite associated cherts is within the upper Black River Group interval associated with the Deicke, Millbrig, and V-7 K-bentonites. Haynes (1992) indicated that the Deicke especially has a dark chert under-layer that helps with its recognition throughout the southern Appalachians. Krekeler and colleagues (1994) indicate that silica is derived from devitrification of volcanic glass as well as silica precipitation from subsurface fluids as they travel through the sediments during multiple phases of diagenesis. They also note, however, that sediments within which the cherts were forming still retained significant porosity thus indicating that silicification at least initiated before substantial compaction took place.

<u>Chert-Rich Intervals of the CBRT</u>

As shown in **figure 9**, there are at least nine fairly widespread chert-bearing intervals in the CBRT.

Chert-rich interval 1

The lowest chert (CR#1) documented is just above the widespread Knox unconformity and is usually composed of angular white chert fragments admixed with a range of dolomite, quartz, and in some cases feldspathic fragments. This assemblage is clearly related to erosion of underlying units during the Sauk-Tippecanoe global sea-level lowstand and is often found best developed in sink holes or karst depressions. This chert accumulation is not found everywhere, but is found in the Ottawa Valley in the vicinity of the Beauharnois Arch, and it is also found in the Lake Champlain region at the base of the Day Point Formation where portions of the Head

Member are polymictic conglomerates with cherts, carbonates, quartzites, and feldspar-rich granitic gneiss clasts. In the southern Appalachians this cherty conglomerate/breccia interval is referred to variably as the Blackford (SW Virginia – East Tennessee), Pond Spring (Tennessee – NE Alabama), Douglas Lake (Sevier Belt-E TN), or Cottingham Creek/Attalla Chert (AL). The chert bed is also found in the Cincinnati Arch region, and in the Upper Mississippi Valley where the basal member of the St. Peter Formation is known to contain chert conglomerates and breccias. In the case of the Upper Mississippi Valley, the unit is referred to as the Readstown Member of the St. Peter.

Interestingly, these angular chert beds have a fairly limited lateral extent and appear to have been deposited in paleokarst depressions cut into the underlying Beekmantown carbonates below the Knox Unconformity (Bridge, 1956; Walker & McLeod, 1991). In the case of the Douglas Lake, the chert is also associated with a very thick (37 m) bed interpreted as a volcanic ash by the previous authors. The combination of extreme karstification (with up to 140' vertical relief), the massive conglomerates and sandstones, and the association of red shales deposited on and within this surface are the basis on which Bridge (1955) postulated the initiation of a major tectonic event. Rodgers (1971), later referred to this as the Douglas Lake Phase of the Taconic Orogeny which he considered the southern equivalent of the Tinmouth Phase (**see figure 2**). Clearly, this chert occurrence is condensed and derived from local erosion and dissolution of pre-existing carbonates during a regional uplift accentuated by the Knox regression.

Chert-rich interval 2

The next overlying chert occurrence (CR #2) is developed near the end of the *B. anserinus* conodont zone. In this case, chert is developed in late transgressive to highstand facies of the Chazy Group and its equivalents. Selleck and MacLean (1988) identify aragonitic fossil debris as being replaced with chert in the coarser grained packstone to grainstone facies of the Day Point to Crown Point interval. It has also been noted as small, 5-8 cm (2-3 inch) nodules within more bioturbated wackestone facies. In most locations, detailed study of Chazy cherts is still necessary to ascertain their specific origin; however, the Crown Point "reefs" are known to contain a diverse assemblage of sponge taxa and argillaceous interbeds are noted for occasional spicules (Welby, 1961). In a few instances cherts are also shown to be associated with thin pyritic and relatively dark-stained hardgrounds where peloids and bryozoan zooecia show evidence of darkened cores suggestive of phosphatic origin.

Thus the appearance of nodular to disseminated cherts in the type Chazy is likely related to early diagenetic processes. Rice (2005) indicates that some of these cherts were clearly diagenetic, but that were likely formed in early diagenesis as there were numerous post-silicification events recorded. Rice provided a post-depositional model wherein tectonism initiated fluid-flow through these limestones, which allowed the silicification to occur. Nonetheless, she was unable to explain the original source of the silica. The proximity of the Champlain Trough to the uplifted Adirondack Arch could provide substantial silica from runoff – as could migration of fluids through subjacent Cambrian Potsdam quartz sandstones during the uplift of portions of the Adirondack/Tazewell Arch during the Tinmouth Phase. Moreover, the association of iron-phosphatic mineralization provides at least some support for an upwelling model for the supply of silica as well.

Outside of the type Chazy region and the Quebec Embayment (**see Ch. 1. Figure 9**), nodular to bedded cherts show an increased abundance in the vicinity of the Pennsylvania Embayment and again south of the Virginia Promontory in the vicinity of the Tennessee Embayment. Within these regions, the Loysburg Formation (central Pennsylvania), the St. Paul

Group (Row Park Formation of southern Pennsylvania), the Poteet Limestone and Elway member of the Lurich Formation (Virginia), and the Lenoir Limestone (Georgia and Alabama) all show chert-rich intervals. In most cases, these limestones host fossiliferous interbeds with excellently preserved fossils that have also been replaced by chert. Although there is some evidence that the southern Appalachian region was also supplied by at least some siliciclastic sediment, there are additional hardground and phosphatic nodule occurrences associated with these strata (Holland & Patzkowsky, 1996) suggestive of phases of nutrient enrichment and sediment starvation along the margin of Laurentia. In addition to cratonic deposits of chert, the lower Mt. Merino (part of the Taconic Allochthon of New York) records an expansion in bedded-cherts coincident with the on-shelf deposits and is likely coeval with CR#2.

Given the substantial uplift and exposure of portions of the carbonate platform during the Knox unconformity and during the ensuing transgression, it is not clear if runoff could have supplied dissolved silica over this wide region. Likewise, if the source of silica was upwelling it is not entirely clear as to how the upwelling would have been generated given the distribution of these cherts and the paleogeographic reconstructions of reentrants. Nonetheless the predominance of cherts in the vicinity of the Sevier Basin and along the flanks of the Tazewell Arch into northern Virginia to southern Pennsylvania suggests a connection with the combined impact of sea-level rise and the initiation of rapid subsidence of portions of eastern Laurentia associated with the first basin-filling phase of the Taconic Orogeny. Substantial accumulation of Chazy strata indicates more subsidence in these areas compared to areas in the vicinity of the promontories. Thus, given the geometries of these embayments and the evidence for rapid subsidence, there is at least the potential that these cherts derived their silica from upwelling sources.

Chert-rich interval 3

The third chert-rich interval (CR #3) appears to be relatively constrained in geographic extent to the region of the Sevier Foreland Basin. The region encompasses the area from the Shenandoah Valley of Virginia, south through eastern Tennessee and into northeastern Georgia. Cherts occur in mass in the upper Lincolnshire Formation (Hogskin Member) as well as in the Rockdell, and uppermost Lenoir of Tennessee. Cherts are also very prevalent in the Rob Camp Limestone of Tennessee. Cherts are also found in the base of the Botetourt Limestone at the base of the Edinburg Formation in north-central Virginia, and in the Effna Limestone of southern Virginia. Age-wise, most of these cherts appear to be coincident with the end of *C. sweeti* conodont biozone and appear just before major dark-shale deposition of the Athens, Blockhouse, Paperville, Rich Valley and Liberty Hall shales found in the Sevier Basin.

Given its narrow distribution (relative to CR #2, and CR # 4), this chert-rich interval is somewhat enigmatic. It appears that this chert occurrence is restricted in some fashion to the Sevier Basin and may be reflective of the foundering of the shelf during initiation of the Blountian Tectophase of the Taconic Orogeny. This event may have dropped the basin to a level where upwelling of nutrient-rich waters was possible. Nonetheless, chert development is minimal once siliciclastic influx in this region occurs. Very few chert intervals occur within the level of the black shales of the Sevier Basin and its carbonate equivalents on the adjacent ramps in this region until well after the first major pulse of coarse-siliciclastics enter the basin.

Chert-rich interval 4

The next overlying chert occurrence (CR #4) is one of the most widespread of all the chert-rich intervals documented herein. Cherts in this interval are constrained to the shallowing phases of this particular succession of strata. In most cases, these cherts are dark grey to black nodules found associated with fine-to-medium grained, calcarenites, or whitish grey chert nodules associated with fine to medium grained micritic facies and vuggy dolomitic intervals. In general the whitish-grey cherts are found in interior regions of the GACB (i.e., New Glarus member of the Pecatonica Formation, L. Gull River of Ontario, L. Camp Nelson of Kentucky, the Ridley of Tennessee, and the Martin Creek of Virginia (Miller & Brosge, 1950; Harris, 1965)), while the darker grey to black chert facies are found associated with the coarser grained bioclastic facies with occasional quartz (windblown?) lamina in the shallowest uppermost facies (i.e. Hostler and Fannettsburg of Pennsylvania and the Wardell and Chatham Hill of Virginia (Brent, 1963)). As such, the lighter colored cherts generally occur in fine-grained, protected facies, and the darker cherts occur in higher porosity sediments deposited under substantially higher energy conditions near the margins of the GACB.

The source of the silica found in these deposits is again not known. There have been a few bentonite occurrences reported from the Fannettsburg (Craig, 1949) and from the Wardell – although they are not consistently identified nor are they named K-bentonites. The restriction of facies at the end of this interval lends little support for upwelling as the source of silica – especially within the more protected facies of the GACB interior – although upwelling could be plausible for the cratonic margin setting provided upwelling continued during regression. Nonetheless depositional facies, within this shallowing succession and again in the next overlying transgressive interval, are much more characteristic of evaporative conditions as evidenced by widespread mud-cracked dolomicrites, evaporite crystal molds, possible chicken wire texture, and stromatolites, especially in Ontario. Combined with the occurrence of frosted quartz grains (wind-blown?) in the Camp Nelson Limestone from the subsurface of Kentucky (Carpenter, 1963), well-rounded (and frosted?) quartz grains in laminae in the upper Hatter of

Pennsylvania (Hostler), "micrite with scattered eolian quartz sand" in the Lower Gull River, and in portions of the Pecatonica of the Upper Mississippi Valley (Lien, 1998), there is evidence for substantial aridity at this time. Moreover, white cherts in the lower Gull River are immediately overlain by beds with gypsum crystals. Thus arid conditions may have allowed for an increase in wind-blown materials as a source of silica into restricted areas of the GACB. For the Upper Mississippi Valley, siliciclastic sediment was definitely supplied primarily from the Canadian Shield (Superior Province and associated regions) to the north (Chetel et al., 2005). For the Cincinnati Arch to Pennsylvania region it is not entirely clear but for the first time in the record of the GACB, there is the possibility that the siliciclastics came from orogenic sources to the east of Laurentia or from locally uplifted areas that became very pronounced at this time along the Adirondack and Tazewell Arches.

Chert-rich interval 5

The next return of chert-rich facies occurs in the late *B. compressa* to early *P. undatus* conodont chronozones (CR # 5) and is coincident with the M4 sequence of Holland & Patzkowsky (1996). This interval also contains at least six to seven named K-bentonites including the well-known Deicke and Millbrig K-bentonites. Cherts are recognized in the Upper Gull River Formation (Moore Hill Member) of Ontario (Liberty, 1969), the House Creek and Amsterdam of New York (Walker, 1973; Fisher, 1977), the Tyrone of Kentucky (Kolata et al., 1996), the Middle Carters of the Nashville Dome (Holland & Patzkowsky, 1996, 1998), the Upper Linden Hall and Upper Mercersburg of Pennsylvania (Craig, 1949), the Lower Eggleston of Virginia, and the Hardy Creek (Harris et al., 1962), middle Moccasin, and middle Bays of Virginia and eastern Tennessee (Haynes & Goggin, 1993). In these cases, the occurrence of cherts appears strongly correlated to the occurrence of K-bentonites including the Deicke and

Millbrig from this immediate interval. In most cases, cherts appear immediately above and below many of these bentonites. Although there are instances of dark blue-black nodular cherts in bioturbated fine to medium grained wackestone to bedded-cherts in more micritic facies not immediately associated with pronounced ash beds. In these instances it is clear that bioturbation was effective in mixing and homogenizing sediments so if thin k-bentonites were deposited, they are not preserved as distinct layers in most of these units.

Therefore, although the source of silica in these cases, appears to be fairly obvious, petrographic analysis of bentonite hosted phenocrysts does not show significant sourcing from volcanic quartz phenocrysts, as most are typically clear and glassy, and do not show obvious evidence for dissolution or chemical etching. Thus if the cherts in this interval are derived from devitrification of volcanic ash post-deposition, the silica is not likely from dissolution of quartz itself but from other components of these ashes possibly prior to deposition. In addition to the volcanogenic devitrification source potential, there have been studies from both modern and ancient settings where volcanic eruptions have been critical in supplying silica for siliceous plankton including radiolarians, diatoms, and some dinoflagellates. As suggested by Ehrlich and colleagues (1996), there is a strong correlation between volcanic ash deposition in marine regions and siliceous plankton blooms. Moreover, Erlich and others (1996) relate volcanic-induced siliceous plankton blooms to periods of expansion (both shallowing and deepening) of the oxygen minimum zone. Such is the case in late Cretaceous oceans where the consumption of oxygen by decomposers is argued to have expanded the oxygen minimum zone thus resulting in increased burial and preservation of organic matter as recorded in the carbon isotopic record. In the pre-diatom Ordovician, it is reasonable to assume that radiolarians and hexactinellid sponges benefited from the incredibly large eruptions of the Deicke and Millbrig (and associated bentonite swarms) and may have proliferated as a result of the widespread volcanic ashes. The coincidence of K-bentonites with subsequent carbon isotopic excursions in this specific interval suggests that a connection is at least plausible for the expansion of the oxygen minimum zone during post-eruptive periods owing to radiolarian blooms.

Chert-rich interval 6

The subsequent K-bentonite-rich interval near the top of the *P. undatus* conodont zone coincides with another interval of widespread chert development (CR #6). Well-developed silicified fossil horizons in the Leray Member of the Chaumont Formation (lower Bobcaygeon) are followed by nodular to marginally bedded black cherts in the bioturbated wackestone to packstone facies of the Watertown Member of the Chaumont Formation. Collectively, these demarcate the next chert-rich interval. Outside of the type region, silicification of fossils and or development of nodular cherts is recorded across much of the GACB during the first substantial transgression of the Tippecanoe Supersequence. Silicified fossils and olive-black cherts have been noted from: the Guttenburg Member of the Decorah Formation, the lower Chandler Falls Formation of Michigan, the Coboconk –Lower Bobcaygeon of Ontario, the Isle LaMotte and Orwell limestones of northeastern New York and Vermont, the Centre Hall and Rodman members of the Nealmont Formation, the Upper Greencastle of the Chambersburg Formation, the Myerstown member of the Jacksonburg Limestone (all of Pennsylvania), the upper Eggleston of Virginia, and the Curdsville Limestone of Kentucky, and adjoining areas. This particular interval is generally easily recognized by the rapidly deepening upward facies pattern capped with the very first shales derived from the Vermontian tectophase. These cherts are also associated with: the recurrence of Chazy-like faunas (stromatoporoids, rugose corals, diverse echinoderms, and a multitude of molluscan forms, including *Maclurites*), bentonites above the level of the Millbrig K-bentonite of the upper Mississippi Valley, and the Cincinnati Arch, which include the Elkport/Capitol and the Dickeyville/Shaker Creek, and immediately precede the first widespread *Prasopora* epibole as described previously.

Despite the widespread chert interval, very little specific research has been done on the Watertown cherts and their equivalents to ascertain their specific petrology much less their diagenetic history or likely silica source. With the plethora of volcanic ash deposits, the abundance of silica producing sponges (including *Brachiospongia*), rapid expansion of radiolarians, and evidence for increasing sea levels and/or increased Vermontian tectonic subsidence, the precise source of silica is unclear. Nonetheless, it is clear that whatever the source(s) of silica, the widespread distribution of these beds must certainly be tied to a significant event (or relatively simultaneous series of events) in the silica cycle. Moreover, with the subsequent Guttenberg Isotopic Carbon Excursion there may be wider implications as well.

Unlike the previous two chert-rich intervals, which represent the widest distribution of cherts throughout CBRT strata, cherty intervals become substantially less widely distributed in the overlying Trenton. Nodular to bedded cherts are usually confined to shallow carbonate-dominated facies and are generally lacking in deeper-water facies where siliciclastics become more than just a minor contributor to the overall sediment load. Occasionally in these deep-water, oxygen-restricted facies, cherts may occur in minor burrow fills or may take the form of minor spicule-rich beds as is the case in some bedding planes in the Martinsburg Formation of Virginia. In some cases, including in the Brannon Member of the Lexington Limestone of Kentucky, cherts appear to be associated with laminated calcilutite facies which occasional become deformed.

Chert-rich interval 7

The relative lack of cherts in the Trenton equivalent interval is particularly intriguing as the supply of siliciclastics is substantial as are the occurrence of K-bentonite horizons that could potentially supply significant dissolved silica. In some cases, abundant clay particles on and under the seafloor may act as an inhibitor for nucleation and growth of cherts (Pfirman & Selleck, 1977; Selleck, 1985). Thus the preservation of chert-rich intervals is not only governed by the availability of silica, but also must be tied to a set of diagenetic parameters (pH, O_2 availability, clay content (absence), etc.) afforded by the conditions of the shallow carbonate platform that are tied to specific events.

Nonetheless, three additional chert-rich intervals are documented in the Trenton and its equivalents. The earliest of which is the most widespread. The lowest post-*P undatus* zone cherty interval (CR #7) occurs in the Kings Falls Limestone of the type region which is defined in adjacent Ontario as the Kirkfieldian Stage. The cherts in the Kings Falls Formation, again take the form of silicification of some fossils as well as minor nodules of dark grey to black chert. These cherts have been noticed in the Kirkfield (upper Bobcaygeon) where they are somewhat larger nodular cherts. Cherts in this interval have also been noted in the coarse-grained facies of the Larrabee Limestone Member of the Glens Falls Formation of eastern New York, as well as in the lower Grier of Kentucky, and in the lower Dunleith Formation of the Upper Mississippi Valley (Beecher-Eagle Point interval). Relatively few chert occurrences have been identified from the Salona of Pennsylvania (although some silicified fossils have been noted) and minor cherty-beds occur adjacent to K-bentonites. The Salona contains an abundance of clay particles that would have prevented nucleation and growth of cherts. Likewise there are cherty and silicified fossil intervals in the Martinsburg Formation, but their position is relatively difficult to place owing to the complex structural deformation of this unit and the lack of a detailed stratigraphic synthesis to date.

Chert-rich interval 8

Overlying the Kirkfieldian aged chert-rich interval, another interval in the Dunleith of the Upper Mississippi Valley is known to produce cherts (CR #8). In this instance, the Mortimer Member shows an abundance of light grey to chalky weathering cherts similar in appearance to those of the intracratonic Black River Group. Occasional silicified fossil occurrences have been reported from this approximate interval but the only additional nodular chert occurrences appear to be in the lower Bigby-Cannon facies of the Nashville region, and a chert-rich interval in the lower-middle Dolly Ridge / Martinsburg of Virginia may coincide with this event. In central Pennsylvania, calcisiltite-dominated, soft-sediment deformed horizons in the Roaring Spring Member of the Salona Formation demonstrate occasional nodular cherts and silicified brachiopods. In this case, dewatering during syn-depositional deformation may have been enough to extrude clay-rich materials thus allowing chert growth in some beds at about this same level.

Chert-rich interval 9

The youngest chert-rich interval (CR #9) appears near the top of the Dunleith in the Wall and Wyota Members. In some localities a cherty breccia has been reported from the base of the Wyota, and may coincide with a cherty intraclastic level in the lower Lindsay of Ontario. Chert-rich intervals, in deformed strata or seismites of the Devils Hollow Limestone of Kentucky, are approximately coeval with these chert occurrences. Interestingly, this interval records resurgence in the occurrence of sponge and stromatoporoid biostromes, as occur in the Strodes

Creek, Stamping Ground, and Devil's Hollow interval of Kentucky, in the middle Catheys Formation of Tennessee, and in the lower-middle Rust Formation of New York. The occurrence of cherts in these latter locations appears to be more sporadic and is often limited to minor nodular occurrences that are often not obvious in outcrop.

Outside of these nine main chert-rich intervals, there have been only a handful of additional cherty intervals recorded in the literature. In all cases, these appear to be rather localized and do not have lateral continuity or lack specific stratigraphic information that allows them to be placed within the constraints of the previously discussed chert-rich intervals of the GACB.

Summary & Implications of Chert-Rich Intervals

Although more research is clearly needed, the occurrence of cherts in the CBRT interval appears to be related to a number of processes. In all cases, the important consideration has to focus on the availability of silica. Given the properties and architecture of the GACB during the Upper Ordovician and the dominance of carbonate production during most of this time, silica availability appears to have been a limiting factor most of the time. Within the CBR interval, when much of the platform was relatively restricted and isolated from open-marine circulation, silica sourcing from marine upwelling is problematic except along cratonic marginal settings. This problem is especially exacerbated during CR #3, when most cherts appear to be associated with a time of significant shallowing and increased restriction on the platform. In this scenario, silica sourcing may have been afforded to the GACB by 1) increased discharge of silica-rich groundwater from adjacent highlands (mostly the Canadian Shield areas) 2) increased delivery of wind-blown quartz-rich silts as might be afforded by a relatively pronounced aridification event, or 3) early wind-blown volcanic ash that may have been precursor to the major volcanic eruptions that are likely to have influenced development of CR#4. The first option would be likely to influence regions nearer the exposed shield and less likely to influence deposition in more distal regions. The latter option, is plausible; however, K-bentonites, although present, are relatively thin and do not seem to be present over widespread areas of the GACB. Thus, it is suggested, in tandem with other sedimentologic evidence, that the GACB likely witnessed significant aridification during the early to mid-Turinian that may have resulted in transport of silica-rich dust to the GACB. This may have been in part due to uplift of the Blount highlands.

As for the chert-rich intervals in the Trenton, most of these deposits appear to have developed in close proximity to the major intracratonic trough (i.e. the Sebree Trough) or the Taconic foredeep. Thus, formation of these chert horizons may have been related to increased rates of upwelling which enhanced the growth of not only radiolarians, but also of sponges that relied on the availability of silica. Likewise, the absence of cherts in these regions at certain times is likely tied to reduction in rates of upwelling or via the introduction of chert-poisoning siliciclastics (clays) that reduced the potential for chert formation (changes in seawater chemistry, sea-level trajectory, availability and abundance of clay particles, etc.). Thus, chert occurrences within foreland basin settings may indeed represent time-restricted facies in the sense of Walliser and are likely governed by a range of unique extrinsic events.

<u>Pulses of Siliciclastics in the CBRT Carbonate Factory & Time Restricted Facies</u>

General Introduction:

Within the carbonate succession of the GACB, shaly or quartz-rich intervals are easily recognized due to their interruption of the otherwise carbonate-dominated Chazy- Black River groups. In some cases, they can be correlated over some distances, and tend to be associated with significant perturbations in the production of carbonate. Commonly these intervals have been interpreted to record sea level fall whereby carbonate production is interrupted by delivery of siliciclastics onto the platform due to evacuation of estuaries and the lowering of base level. In some instances, these mixed siliciclastic-carbonate intervals become substantially more fossiliferous than associated carbonates and thus have been interpreted to represent transgression over carbonates. Alternatively, in some instances, these siliciclastic-rich intervals may be tied directly to climatic oscillations whereby sediment could either be transported by wind during aridity events (quartz-dominated associations with eolian characteristics) or by fluvial transport during humid climatic events (silty-shale-dominated or polymictic conglomerates).

Within the CBRT and especially within the Black River Group, where limestones are relatively pure $CaCO_3$ (commonly 95% or better), the occurrence of siliciclastic deposits clearly represents a perturbation in the carbonate factory. Moreover given that these siliciclastic-rich intervals do have considerable lateral continuity, although they are not lithologically homogeneous from location to location, they can be recognized by the increase in siliciclastics as they can make up as much as 50% or more of the succession in which they occur. For instance, the Glenwood Shale of the upper Mississippi Valley has been correlated widely in the subsurface and is considered the equivalent of the Wells Creek in the subsurface of Ohio, Kentucky, and elsewhere. The Glenwood also coincides with at least part of the siliciclastics in the Sevier Basin of Tennessee. In another instance, the siliciclastics of the Spechts Ferry Shale (Lower Decorah) are well correlated in the upper Mississippi Valley (Iowa, Minnesota) and have been correlated historically with shaly-carbonate units in Ontario and New York (Kay 1931).

In all cases, the origin of the siliciclastics units has been poorly constrained. Nonetheless, recent data for the siliciclastics in the Decorah of Iowa show evidence of significant shifts in provenance during sea-level events where sediment sources shift back and forth between local and more regional sources. Specifically, as supported by neodymium isotope data presented by Fanton and colleagues (2002), Spechts Ferry shales suggest local provenance from the Transcontinental Arch area during shallowing conditions. In contrast, upper Decorah Guttenberg and Ion interval siliciclastics indicate a contribution for the first time from Taconic sources far to the east.

Thus, it appears possible that some of these siliciclastic events have the potential for widespread development and might record intervals of sea-level change (as is the conclusion drawn by the Fanton and colleague study). Nonetheless, during deposition of most of the Black River the magnitude of water depth change is presumed to be significantly less than the terminal Black River flooding event discussed by Fanton and colleagues (2002). Moreover, most siliciclastic events on the platform (in the Chazy-Black River) appear to coincide with late regressive basin-filled phases, which would tend to favor more localized as opposed to regional sourcing of sediments. Furthermore, during much of this time regional terrane was apparently very subdued, and therefore the supply of siliciclastics onto the GACB was probably limited to a few areas. Sediment sources may have included some areas of the GACB near the Transcontinental Arch and along the Canadian Shield along the northern margin of the Michigan Basin and southern Ontario along the exposed Grenville Shield. Other sediment sources can also be attributed to areas nearest the tectonically active Blount highlands or potentially from the Ozark Dome.

Alternatively climatic processes may also have played a role. In the sub-tropical belt, it is plausible for desertification to occur resulting in wind-blown sediment deposits. As known from modern settings, aridity events are capable of transporting sand, silt, and dust-sized materials over wider areas. Likewise, under episodically humid or monsoonal conditions uncharacteristically high riverine discharge may also transport significant volumes of sediments into marine settings where they can be redistributed over wide regions – especially during regression. In either case, the delivery of siliciclastics and associated nutrient-enrichment is likely to have a significant impact on the carbonate factory on at least local scales – if not more regional scales.

Nonetheless, within the constraints of biostratigraphic zonation and previous stratigraphic studies, it is readily apparent that several siliciclastic-rich zones can be identified and used for correlation at least within local sub-basins. Moreover, as suggested here there is also the potential that some of these siliciclastic-rich intervals coincide with coeval siliciclastic deposits in other sub-basins of the GACB even if they are not directly correlative with one another. As presented here, the occurrence of relatively synchronous siliciclastic-rich intervals from different sub-basins suggests the need for further provenance studies. In the meantime, the following discussion focuses on seven main siliciclastic events that initiated after deposition of the basal transgressive quartz sandstones of the St. Peter Group and its equivalents leading up to and including the first siliciclastics of the Trenton Group.

Collectively, the first six siliciclastic-rich intervals are most-commonly recognized from platform successions. In weathered outcrop exposures they tend to be recessive shales with a greenish coloration. Commonly, these intervals are dominated by recessively-weathering, green platy shales interbedded with variable amounts of thin ribbon micrites, nodular lime mudstone, or dolomitic ribbon limestones. Alternative variations occur where the greenish coloration is associated with massive beds containing minor shales, substantially more quartz, and carbonate-usually dolomitic limestone or dolostone (**figure 11**). In unweathered sections and cores, the color can be a somewhat duller greenish grey that is often not as easily identified as being siliciclastic-rich unless carefully investigated. This is especially true when the ribbon micrite interbeds compose more than a few percent of the interval. These beds are not to be confused with the metabentonites commonly recognized including the Deicke K-bentonite that also has a very distinctive green color (Kolata, et al., 1996). The former intervals usually show no evidence of swelling clays common for most altered volcanic ash beds.

In most cases the source of the distinct green coloration has been shown to be somewhat difficult to pinpoint as petrographic studies have been limited on these specific interbeds. In many cases the greenish color is attributed to the presence of glauconite in either detrital or authigenic form, as undifferentiated glauco-chlorite lithologies, or simply as greenish argillaceous dolomitic limestones (Textoris, 1968; Melchin et al., 1994, Dever et al., 1994). In this case, the greenish color is not identified to source. The former mineral group (glauconites) is highly diverse in its composition and can be associated with a large range of mineral forms. Recent studies of green clay minerals suggests that the only common denominator of "green clay minerals" is that they all contain iron in both valence states – that is reduced (Fe^{+2}) and oxidized (Fe^{+3}) iron (Velde, 2003). In most instances these green clay minerals are associated with a range of other clays and mixed layered clays (i.e. illite, smectite, chlorite), but also can be characteristic of quartz-rich carbonate intervals, especially in cases where micro-fossils and pellets become the site of green mineral formation (Kronen & Glenn, 2000; Velde, 2003).

Figure 11: Representative examples of siliciclastic-rich intervals interbedded with massive, cyclically-bedded carbonates characteristic of the Chazy-Black River interval. The green marker bed of the Lake Simcoe District is the same bed as the Pittsburg Quarry bed of Conkin (1991). As indicated by Cornell (2001a), the upper green marker bed can be correlated over a

distance of at least 250 kilometers or more. A very similar interval has also been described from sections in Manitoulin Island (northern Lake Huron). If this is indeed the same stratigraphic interval it would extend the distance to some 550 km. Also shown is a representative green shale interval from near the top of the Tyrone Formation of Kentucky termed informally the "Marcellus shale" for exposures southwest of Marcellus, Garrard County, Kentucky.

Given the requirements for mixed Fe sources, the formation of glauconite minerals has proven to be enigmatic in most stratigraphic successions. In the rock record, it has most commonly been associated with transgressive systems tracts and relative sediment starvation in off-shore deep-water shelf and slope settings (see Udgata & Savrda, 2007). In these instances the sediment-water interface and the redox-boundary is generally held stationary for a period of time which allows for migration of pore fluids and the increased incidence of reworking and winnowing of the sea-bottom (Kronen & Glenn, 2000). Collectively these processes can provide the necessary "mixed valence" conditions necessary for glauconite formation (Velde, 2003) provided the seafloor is not fully cemented by early diagenesis (hence reducing porosity) and that there is an Fe source (typically siliciclastic rocks on the continental shelf). Under these conditions, it is possible that glauconite formation may enhance cementation and stabilization of seafloor substrates and help in the process of hardground formation.

In contrast, shallow water siliciclastic environments today show increased incidences of glauconite precipitation coincident with increased fertilization from nutrient-rich runoff (Nelson et al, 1994). In these cases, nutrient-enrichment is shown to facilitate algal blooms and localized anoxia near the seafloor thus allowing increased Fe^{+2} concentrations. Reworking of these sediments by storms or biologic activity is likely to establish the necessary "mixed valence" conditions necessary for glauconite formation. In this case, glauconite can form if Fe supply is substantial with respect to sulfate concentration. Under these conditions, glauconites can form on the interior of aragonite/calcite shell materials, or as coatings on detrital particles including quartz and a variety of coarse mica minerals, or within fecal pellets or carbonate intraclasts. This scenario for glauconite formation is particularly significant because in most stratigraphic syntheses, glauconite is thought to form predominantly during transgressive phases, and this interpretation could be in error. Clearly as documented in this case, the formation of glauconite can occur in shallow water environments and can be unrelated to sea-level rise or the associated sediment starvation, and stabilization of the redox boundary.

In the Ordovician there are several instances where shallow-water depositional environments, and even peritidal settings, are inferred for glauconite deposition– especially in the Cambro-Ordovician of the southwestern United States (Chavetz and Reid, 2000). In this scenario, glauconites formed in variable coarse-grained, shallow sub-tidal to tidal flat settings where they are thought to have precipitated directly onto or just beneath the seafloor. This setting is more similar to nutrient-enriched, higher energy, shallow-water deposits as described previously, than they are to deeper-water shelf to slope deposits as are usually inferred for glauconites.

Moreover, in the eastern U.S. there is at least one study documenting glauconite formation that appears to have formed during regression. In this case, glauconite-rich quartz sandstones have been recognized in the St. Lawrence Platform of Quebec. The succession has been recently re-interpreted to represent regressive conditions preceding a widespread unconformity on the Laurentian shelf (Longuepee, 2005; Longuepee & Cousineau, 2005). In this study, the regressive sequence in the Anse Maranda Formation has non-typical aluminum-rich glauconites within massively-bedded, shallowing-up, bioturbated sandstones with some detrital clay matrix. Glauconite is not present in the well-winnowed, clean sandstones immediately below. The underlying clean sandstones appear to exhibit deepening-up or transgressive signatures and were cemented by early diagenetic carbonates. Although glauconite is known to form in similar settings today, the glauconites of the Anse Maranda are not adequately assigned to the transgressive model. Instead, the predominance of intense bioturbation in thick sandstones suggests increased organic detritus, which may have resulted from high rates of productivity on the shallowing shelf. If the evacuation of nutrient-rich sediments from estuaries and rivers during regressive events was substantial enough to generate periodic anoxia, glauconite formation in this setting may be more similar to glauconites forming in the Gulf Coast region today as reported by Nelson and colleagues (1994). This alternative nutrient-evacuation model could also explain the source of the highly weathered accessory clay minerals within the otherwise quartz-rich sandstones and may explain the Al-rich glauconite.

Thus the presence of glauconite-rich intervals in the stratigraphic record of passive continental margins needs to be carefully considered. Moreover, their occurrence depends on several factors including: 1) a requisite iron source (weathering of Fe-rich minerals), 2) a mechanism to fluctuate redox conditions so that reduced Fe can be produced and mixed with oxidized Fe, and 3) a sink for the reduced Fe on or just under the sea floor – in most cases this can be another silicate mineral, a sulfide, or any carbonate mineral phase (Velde, 2003). In the case of the glauconite-rich intervals of the Upper Ordovician discussed here, the occurrence of characteristically green, clay-rich, siliciclastic facies is in need of investigation as the origin of many of these horizons is not yet understood. The following discussion focuses on the relative position of these intriguing intervals.

Siliciclastic Events in CBRT:

<u>Siliciclastic Event 1</u>

The lowest siliciclastic event (SE-1) is perhaps the least well-constrained of all of the siliciclastic-rich intervals. This lowest event is contained within the St. Peter sandstone of Iowa and is interpreted to coincide with a shale-rich interval in the Crown Point Formation of New York. Recent investigations of the St. Peter Sandstone of the Decorah, Iowa region reveal the presence of distinct laminated greenish -dark gray shales just below the Tonti Member of the St. Peter. Tentatively termed the "Freeport Bridge Shale" by Young and colleagues (2005), this interval has yielded a restricted, yet well-preserved, fauna including "Chazyan-aged" conodonts, lingulid brachipods, and a number of macrophytic algae, and crustaceans (Liu et al., 2006). Based on conodont assemblages and lithologic similarities, this same interval is roughly time equivalent to the hard platy shales of the Geiser Quarry Member of the Dutchtown Formation from the southwestern margin of the Illinois Basin in Missouri. This same interval is also recognized from the subsurface of Kansas where the upper member of the St. Peter is underlain by beds of light to dark-green shales, mixed with sandy mudstones and some sandstone. It is easily recognizable in northeastern Kansas but becomes more prominent to the south and west toward the Oklahoma border (Goebel, 1968). Nadon and colleagues (2000) investigated this interval within the Michigan Basin and recognized several small-scale sequences (on the order of a few meters in thickness) showing very similar lithologies suggesting that this interval might have a wider distribution than assumed by J.M. Young and colleagues (2005).

In the western-most regions, sediment is likely derived from the Transcontinental Arch and the Ozark Uplift respectively- while sediments in the Michigan Basin likely originated from the Canadian Shield. Nonetheless, the Freeport Bridge Shale, and the Winneshiek Lagerstätte it contains, is apparently limited to a small localized region where it ranges from less than 3 meters to nearly 38 meters in its only known outcrop. If it was distributed over a wider area, it may have been removed by subsequent erosion prior to deposition of the Tonti Member of the St. Peter.

A similar dark-greenish gray shale unit occurs within the lower Crown Point Formation from the type Chazyan area. In a core from Plattsburg, New York, the shale-unit stands in stark contrast to typical Crown Point grainstone, packstone, and fossiliferous wackestone facies. This shale dominated interval is uncharacteristically barren of fossils compared to the rest of the Crown Point although it does contain small ostracod and bryozoan fragments on some bedding planes. Equivalent dolomitic-silty shales have also been noted from the Hog's Back/Laval formations of the Ottawa Valley/St. Lawrence lowland region (Salad Hersi & Dix, 1997), – and are noted to show evidence for truncation beneath overlying units within individual outcrops. The shales in the Lake Champlain to St. Lawrence lowlands were likely sourced from the nearby Adirondack Arch, the developing Beauharnois Arch, and or the Grenville Shield. Paleocurrent analyses in southern Quebec, suggest a southwest directed flow. Thus these shales may have been sourced from the Grenville Shield and transported into the Lake Champlain Trough region from the northeast.

In Pennsylvania, carbonates of the Milroy Member of the Loysburg Formation show very little evidence for a shale-rich interval and are characteristically pure limestones and dolostones. However, the uppermost part of the Row Park Formation in the Cumberland Valley, although dominantly massive carbonate, shows a significantly higher percentage of silt-sized quartz grains relative to underlying or overlying units (MacLachlan, 1967). The silty-interval occurs above the first occurrence of *Maclurites magnus* and thus sits in approximately the same position as the shaly-interval in the Crown Point Formation of New York and may be equivalent. However the source of the silt-sized quartz component is enigmatic. These grains may have been transported some distance either as eolian suspension or by marine currents. Thus as defined, SE-1 appears to be represented in at least a couple of outcrop regions near clearly defined, but disparate, sediment sources (relict pre-Taconic structures). Nonetheless, their occurrence represents a fairly significant change in facies compared to surrounding rock types.

Siliciclastic Event 2

Superjacent to the St. Peter sandstone proper, in the Illinois Basin is an interval of green shale approximately 2 meters in thickness (Templeton & Willman, 1952). This platy shale has been recognized as the uppermost member of the Ancell Group and was named the Harmony Hill Shale member of the Glenwood Formation and represents SE-2 in this study. This same shale is thought to correlate with the uppermost member of the St. Peter in the Illinois Basin where the equivalent is referred to as the Starved Rock Member. Fraser (1976) describes the Harmony Hill as a bioturbated to fissile shale adjacent to but landward of the sandy Starved Rock. He interpreted these as a back-bay, barrier island complex deposited during a single shallowing event after deposition of the Tonti Member of the St. Peter. Both sit immediately below the Chana Member of the Pecatonica Formation or what have been called the Daysville-Hennepin Members of the Glenwood Formation (Fraser, 1976). The basal Daysville is composed of a brecciated to conglomeratic dolomitic calcisiltite facies that becomes dominated by carbonates typical of the Pecatonica Formation above.

Further south in the Illinois Basin, the lateral equivalent of this unit is the Augusta Member of the Joachim Dolomite of the Ancell Group. Locally, the base of the Augusta contains an interval of green shales and siltstones that immediately overlie a very widespread "St. Peter type" sand (Templeton & Willman, 1952). These former authors defined this sand as the widespread Tonti Member of the St. Peter - a combined eolian – marine sand interval.

Again, in both outcrop and in core, the Harmony Hill interval appears to have been truncated in at least some areas.

Nonetheless, the Glenwood has been correlated in the subsurface, using the St. Peter and the relatively high gamma-ray signature of the shales as a guide, eastward across the Illinois Basin into Kentucky and Ohio where it is correlated with the Wells Creek Formation (Stith, 1979). In this region, the quartz-dominated St. Peter and the overlying green, shale-rich, Wells Creek are not preserved in the area of inferred topographic highs. Moreover, these units are preserved in the vicinity of topographic low areas where the shales are often interbedded with dolostones and dolomitic limestones. In these cores, the waxy, dolomitic, pyritic green shales, and argillaceous limestones and dolostones are dominant, but they are occasionally interbedded with minor brown to black shale (algae-rich) stringers with some associated fine-sandstone and siltstone (Wickstrom et al., 1992). The coarser lithologies are more prominent above the contact with the overlying Camp Nelson Limestone, which has been found to contain frosted quartz grains (Carpenter, 1965). Stith (1979) refers to this zone as the "lower argillaceous" interval of the Black River Group – but its proximity to the Wells Creek predates type Black River.

In the type Chazy region the equivalent unit, in the vicinity of the *C. sweeti – P. aculeata* conodont zone, shows a transition from reefal facies characteristic of the underlying Crown Point into dark-gray, silt-laminated calcarenites interbedded with greenish argillaceous calcisiltites of the Hero Member of the Valcour Formation. This interval is quite fossiliferous with sponge biostromes, and numerous ostracods as well as "*Solenopora*" red algal nodules. Recent reports of type *Solenopora* suggest that some forms assigned to this genus may not be algae but may be chaetetid sponges (Riding, 2004), nonetheless, toward the Adirondack Arch, this interval becomes more dolomitic with carbonates containing somewhat larger quartz-grains interbedded with partings of brown to black shales. Near the top of the succession in the Plattsburgh Quarry, thin quartz-pebble beds have been reported from discontinuous beds. The abundance of quartz-grains, pebbles, and even larger intraclasts of a variety of carbonate and metamorphic/igneous lithologies increases substantially into the overlying upper Valcour Formation. The increase in grain-size in this interval and in the equivalent Hog's Back of Ontario has been interpreted to represent significant uplift and erosion along the Beauharnois Arch and perhaps even the Adirondack/Canajoharie Arch (Salad Hersi & Dix, 1999; Salad Hersi and Lavoie, 2000) at this time. It may also represent a coincident sea-level drop made more pronounced in some areas.

In southwestern Ontario, in the Lake Simcoe District it is not clear if the basal Shadow Lake maybe equivalent with SE #2. The Shadow Lake Formation, and its eastward equivalent – the Rideau Formation of the Kingston, Ontario region, are composed almost entirely of siliciclastics and minor carbonate interbeds especially near the top. It ranges up to ~14 m thick and is altogether missing in other areas where it is overlapped and truncated by subsequent deposits. The interval rests directly on Precambrian Grenville basement and is dominated by red and green shales, interbedded with arkosic to quartz sandstones, and conglomerates – with clasts up to 8 m in diameter (Okulitch, 1939; Liberty, 1969). It is entirely possible that the lower Shadow Lake may in part be equivalent to the Wells Creek-Glenwood (SE -2) shale-rich interval as reported by Wickstrom and colleagues (1992). As reported by Liberty (1969), the shales are often calcareous and may contain frosted quartz grains and pebbles that are characteristic of portions of the Harmony Hill – upper Wells Creek. This unit, like the Wells Creek – Glenwood is widespread in the subsurface of southern Ontario and often contains clasts of underlying Cambro-Ordovician units. Without any noticeable fossils, Liberty considers it to represent very shallow-water "deltaic" type deposits that extended from the Canadian Shield south into the

Appalachian Basin and westward beyond the Algonquin Arch into the Michigan Basin. It may also be included in the uppermost "sequences" identified in the Michigan Basin (Nadon, et al., 2000).

In New York State, the basal Pamelia Formation everywhere overlies a distinct green coarse to fine-grained shaly interval recognized by early workings including Cushing and colleagues (1910). This interval is present in the type Pamelia region as well as in the mid-Black River Valley including sections near Lowville. It retains its characteristic green-siliciclastic signature but also picks up some calcareous interbeds and basal conglomerates as in Ontario. In well-weathered outcrops, the green color is replaced by buff to red oxide coated surfaces that indicate the presence of iron in the rocks. These beds have always been included and mapped along with the Pamelia Formation, but they are distinctly set off from the typical sandy-dolomitic carbonates that designate the Pamelia. In appearance they are nearly identical to facies seen in the Shadow Lake and are likely its eastward equivalent – deposited along the flank of the emergent Adirondack Arch.

In the subsurface of northwestern Pennsylvania, Berg (1980) recognizes the Shadow Lake equivalent but it is not expressed at the surface in Pennsylvania, and he does not define its exact lithology. In this region, it does not extend across the Rome Trough. In central Pennsylvania, SE-2 is not clearly represented or developed as a shale-rich interval. In this region, facies are clearly dominated by carbonates and only occasional interbeds exhibit any shale component. However, the uppermost member of the Loysburg (the Clover) contains fine-grained, wavy-laminated ribbon limestones interbedded with argillaceous seams containing ostracods and abundant *Girvanella* nodules (Swain, 1957). This interval is underlain and overlain by substantially more fossiliferous and more massive carbonates. Due to the argillaceous component in this interval, compared to the nearly pure carbonates, the Clover is usually recognized on wireline logs, but it is not confidently traced into New York or Ohio (Fettke, 1948) where it could be related to the Wells Creek.

Along the cratonic margin in the southern Appalachians, prior to the end of the *N. gracilis* graptolite zone, dark shale (Paperville, Athens, Blockhouse and equivalents) deposition was well-underway in the Blountian Basin (Finney et al., 1996). Thus in this region, SE -2 is slightly predated by the initiation of black shale deposition in the Sevier Basin. The first coarse-siliciclastics and mega-conglomerates appear in this region after black shale deposition but prior to the end of the *C. sweeti* conodont zone. This first pulse of coarser clastics (early flysch phase) is admixed with shales and carbonates referred to as the Tellico Formation. The unit is known from Virginia and Tennessee through Georgia where it is predominantly coarse-grained, iron-rich sandstone with interbedded shales. Butts and Gildersleeve (1948) recognize lenticular conglomerate bodies at the base of the unit in Georgia. In Tennessee where it is substantially thicker, the Tellico is represented by calcareous shales interbedded with medium to coarse grained lenticular sandstones (Neuman, 1955). In southwestern Virginia, Cooper and Cashion (1970) also recognize a relatively thin interval of sandstones equivalent to the Tellico above the Paperville Shale and in facies similar to the Liberty Hall Formation. The Tellico is clearly a progradational shallowing facies with sediments derived from the newly uplifted Blountian highlands to the southeast of the basin. In this stratigraphic position, it is approximately equivalent in time to shale-rich Wells Creek – Harmony Hill deposits on the GACB.

Siliciclastic Event 3

Siliciclastic interval three (SE-3) occurs a few feet above the base of the Gull River Formation in Lake Simcoe area of Ontario as described by Liberty (1969) and at the top of the

Shadow Lake as defined by Okulitch (1939) (**Figure 12**). SE #3 was considered the uppermost part of the Shadow Lake by Okulitch (1939) as he originally described the unit to include all impure siliciclastic-rich beds. Liberty demarcated the boundary on the basis of the level of the first carbonates which occurs below the uppermost terrigenous-influenced beds. In either case, the lower green marker bed as it is referred to represents SE #3 and is usually no more than 1 meter thick in the Ontario region. This particular interval is easily recognized on the basis of its characteristic green coloration – but the unit tends to be more massive and shale-poor. It is often considered a bioturbated dolostone with terrigenous clasts, sand grains, and minor disseminated clays. It often displays an intraformational conglomerate at its top which contains pyrite nodules and reworked glauconitic intraclasts and peloids in some localities (El Gadi, 2001). Where cements are not as well-developed, the unit weathers back as a soft recessive interval with abundant medium to coarse sand grains in an argillaceous fabric.

This unit is immediately overlain by a 1-2 meter thick limestone-interval rich in fossiliferous beds rich in ostracods, a few brachiopods, medium-sized gastropods, and some pelmatozoan forms – this was the base of Okulitch's Gull River. The unit has also been reported to be rich in *Solenopora* nodules in some locations. Interestingly in several localities mollusk fossils exhibit a very distinct glauconite lining on the interior of shells – or make up the entire fossil where preservation is moldic. In a few instances the lumens of pelmatozoan columnals have also been observed to contain glauconitic linings. The association of these more fossiliferous limestones immediately above the massive green bed (and the distinct glauconitic beds) allows for regional correlation in the Lake Simcoe region. A similar succession of beds is found in the abandoned Marmoraton Mine at Marmora, Ontario. In this setting, it is very easy to see the full succession of Shadow Lake from the Precambrian through the Coboconk. The

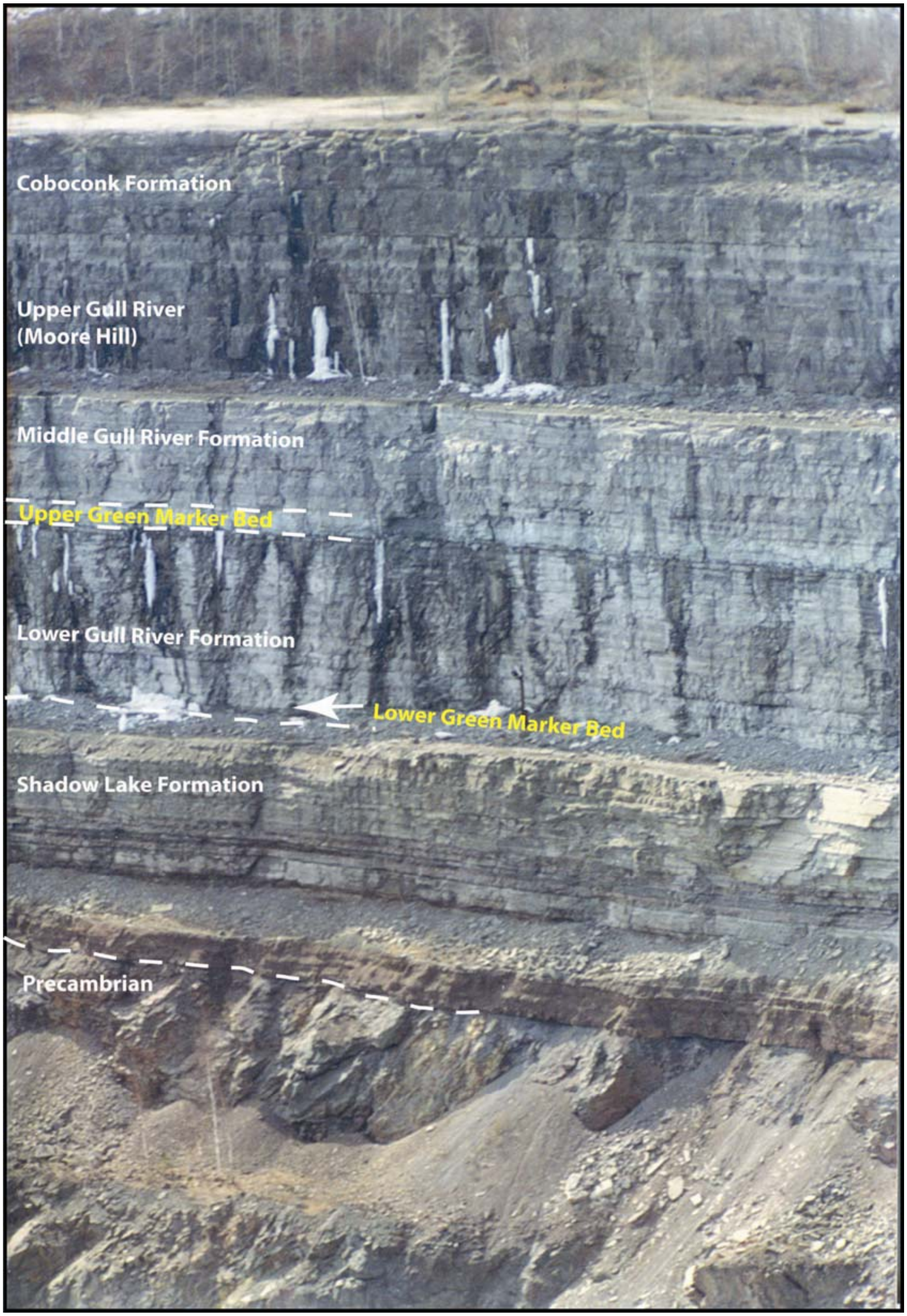


Figure 12: Marmoraton Mine, Marmora Ontario. Gull River Group and overlying Bobcaygeon/Coboconk are shown relative to the position of the upper and lower green marker beds as described herein.

Shadow Lake is particularly thick in this area and the overlying Gull River interval becomes very cyclic in its overall appearance as seen in **figure 12**. However, the green-glauconitic fossil bed has not yet been located nor has it been located explicitly to the east in the Kingston region.

Nevertheless, in the region between Kingston, Ontario and the type region in New York State, the basal Pamelia Formation is set off from the Shadow Lake/Rideau equivalents by up to 2-3 meters of dark grey to black micritic wackestones. They lack evidence for substantial bioturbation but they are often fossiliferous with corals, brachiopods, and very large cephalopods. These are the equivalent of the lowest Gull River beds recognized by Liberty (1969) and are capped with ~1.5 m of somewhat sandy, buff-weathering dolomitic limestones and interbedded shales. Because of the shales, these beds often weather back substantially and are only exposed in a few localities. Unlike the equivalent lower green marker interval in the Lake Simcoe Region, these shales are not the characteristic green but weather to a yellow-brown color and are fairly dark brown on fresh surfaces and are similar to organic-rich (algae-rich) kukersite shales. They contain a more restricted fauna (compared to the underlying black limestones) and exhibit mudcracks on some surfaces. These shaly beds are immediately capped by a massive-bedded intraclastic sandy dolostone that is overlain by an upward deepening succession of fossiliferous black limestones (often weathering buff) with occasional stromatolites and diffuse crinoid occurrences. In a few areas, especially in the southern Black River Valley, this interval also contains a few diffuse oomicrites (Textoris, 1968). These beds often show the first evidence of significant bioturbation and are likely the tie into the bioturbated or burrow-mottled bed of Lake Simcoe workers (Liberty, 1969; Grimwood et al., 1999; El Gadi, 2001). Thus constrained these shales represent a fairly widespread flux of siliciclastic sediments after an initial flooding event and prior to a large deepening phase leading into the deposition of typical Black River Group lithologies.

Outside of the Ontario-New York region, SE-3 likely corresponds to the "upper argillaceous" interval recognized in the subsurface of Ohio (Stith, 1979). It is very easily recognized at the top of the "Carntown" lithographic limestones both using wireline logs and in cores. The upper argillaceous interval lies immediately above an interval of white chert nodules and is associated with dolomitic limestones and interbedded dolostones. The characteristic green color is typical for the interbedded shales and is similar in appearance to the green marker bed interval of the Lake Simcoe region of Ontario. The argillaceous component becomes less distinct to the south into Kentucky but is still recognizable in cores where a few seams of green shale partings are found above the white chert horizons. In central Kentucky, outcrops of the High Bridge Group expose a 1.5-2.0 meter-thick recessive weathering green shale below the mottled beds recognized by Stith (1979) and approximately 8-9 meters below the middle "white marker bed" recognized by Ettensohn (1992). The bed is immediately overlain by burrowed dolomitic limestone and intraclastic breccia and grainstone bed reminiscent of the lower Gull River. Ooids have not been reported from this interval.

Details of this particular interval in the Upper Mississippi Valley - Illinois Basin region are still somewhat unclear. However an argillaceous and dolomitic interval occurs in the uppermost Pecatonica Formation of the Platteville Group. In this region, Kolata and Noger (1991) were able to differentiate upper Pecatonica from the overlying Mifflin (McGregor Formation) on the basis of gamma-ray logs. The slightly higher gamma-ray readings coincide with the approximate position of the New Glarus Member of the Pecatonica. Mapping by Nelson (1996) shows the Pecatonica to be dominated by intercalated dolomite and limestones with thin shale interbeds. The Dane and New Glarus members of the Pecatonica contain chert in the vicinity of an interval of brown-laminated dolomite and medium gray shales. The Pecatonica including the New Glarus interval is shown to be truncated or all together absent at the base of the McGregor in some areas outside of Iowa (Brandt, 1978; Witzke & Bunker, 1996; Ludvigson et al., 2004). Nonetheless a very thin iron-rich dolomite demarcated with a “ferruginous hardground” sits just below the base of the fossiliferous and oolitic Brickeys Limestone Member of the McGregor Formation. It is likely the approximate position of the shaly interval. It may be this ferruginous interval that is recognizable on wireline logs.

As mentioned previously, the green marker interval (SE-3) in the Gull River and Pamelia Formations of the Ontario-New York region is typically overlain by a sandy dolomitic, and often polymictic intraclast bed. This bed is clearly indicative of substantial erosion prior to deposition of the main phase of the Black River Group. A nearly identical bed is recognized in the Lake Champlain region and is usually considered to be the Chazy-Black River unconformity – but little evidence exists for the greenish shaly interval. In contrast, in central Pennsylvania, a dark, tan-weathering interval at the top of the Hostler Member of the Hatter Formation is capped by oolitic - intraclastic beds of the Snyder Member of the Benner Formation (Kay, 1943; Faill et al., 1989). The dark-tan weathering interval is sparsely fossiliferous and laminated but contains chert and silicified fossils. In outcrops near Union Furnace, this interval also contains thin (3-4 cm) thick bioturbated black shales interbedded with a few coarser-grained grainstones just below the oolitic grainstone that represents the base of the Snyder. These shales often contain black organic-rich filamentous strands characteristic of non-calcareous macrophytic algae (LoDuca, 1995). Although this interval does not contain appreciable dolostones or siliciclastic sandstones – the succession of silicified fossiliferous limestones with nodular cherts, and interbedded black-shales capped by oolitic intraclastic beds is at least reminiscent of the succession elsewhere. However, without a nearby source for coarser-grained siliciclastics the interval is not exactly similar. Nonetheless, herein it is inferred that the black shales in the uppermost part of the Hatter Formation are the equivalent of SE-3.

At the time SE-3 was deposited on the GACB, sections in the vicinity of the Sevier Basin were continuously influenced by siliciclastic deposition for a period prior to initiation of deposition of SE-3 on the platform. In eastern Tennessee, the Ottosee Shale and associated sandstones had already accumulated to thicknesses of up to 600 meters. They are typically brown to brownish gray on fresh surfaces and weather to yellowish-brown in exposed outcrops (Milici, 1973). Commonly the Ottosee is interbedded with coarse iron-rich limestones and pink encrinites. The formation is divided into an upper and lower unit on the basis of a widespread sandstone body that occurs near the top of the interval. The Bacon Bend Member was documented by Neuman (1955) to exhibit submarine soft-sediment "slump structures" within an interval of calcareous sandstones below the top of the Sevier and below the Bays Formation. Coarse-grained limestones below the Bacon Bend contain occasional chert nodules and silicified fossils.

On the northwestern margin of the Sevier Basin, in the approximate stratigraphic position of the Bacon Bend, the middle-upper Ottosee has been reported to contain an interval of uniquely preserved ooids from a 10 meter-thick interval in the area of Knoxville, Tennessee (Cantrell & Walker, 1985; Weaver, 1992). The unit contains small patch reefs and several ooid-rich lithotypes. Cantrell and Walker (1985) indicated that at least for some of the ooid intervals, petrographic analysis of cements shows evidence for initial marine diagenesis followed by meteoric cements suggesting that they may have been deposited during sea-level lowstand.

Moreover and most importantly, Weaver (1992), recognized that many sedimentary particles show evidence for authigenic phyllosilicates (physils). These occur within calcite ooids, pellets, bioclasts including echinoderm pieces, laminated algal mats, and on the surfaces of large quartz grains that may have been coated with cyanobacteria. These physils were identified as green Fe-chlorites, Mg-chlorites, corrensite and other swelling clays and were not found in the cements, which Weaver suggests is an indication that they were emplaced before significant burial diagenesis. The apparent structure of some of these physils suggests that they may also have been mediated by various microfauna near the seafloor under micro-reducing conditions. The occurrence of this unique ooid interval is particularly important because it coincides roughly with the ooid-rich interval in the Snyder, Pamelia, Lower Gull River, and Brickeys. It is also important because of the occurrence of similar green mineralized coatings on mollusk, and echinoderm debris in the lower Gull River Formation of Ontario. Collectively these data suggest the potential for similar processes acting under the same conditions and maybe even at the same time due to an as yet undetermined event.

Siliciclastic Event 4

The fourth siliciclastic interval on the GACB (SE-4) has been referred to in the Lake Simcoe region of southern Ontario as the "Green Marker Bed" (Grimwood et al., 1999; El Gadi, 2001). This marker bed is located at the top of the Lower Gull River Formation (as opposed to near the base of the Gull River as with SE-3) as interpreted by Grimwood and colleagues (1999) as well as El Gadi (2001). This bed, herein, referred to as the "upper green marker bed," lies near the base of Liberty's (1969) Middle Gull River (**see figure 11 and 12**). The bed is located above a fossiliferous interval containing distinct, medium-bedded, burrowed wackestones, referred to as the burrow-mottled bed or bioturbated bed (Grimwood et al., 1999). The green marker bed of these former workers is actually the uppermost of two fairly widespread green marker intervals. It is composed of medium to light green, argillaceous, fine-grained microcrystalline dolostone with siliciclastic silt and vugs and disseminated glauconite and pyrite nodules. Petrographic analysis by Grimwood and colleagues (1999) indicate that this unit is associated with ferroan dolomite (likely after primary dolomite), as well as gypsum, anhydrite, and celestite minerals (sometimes replaced by calcite) or in weathered outcrops they dissolve and produced molds of the bladed crystals suggesting no replacement occurred at all. The unit also contains a terrigenous fraction that includes sand sized quartz, and feldspar grains as well as clay minerals. Collectively, the bed contains up to 25% siliciclastics. The upper green marker bed sits immediately above a coarse-grained dolostone bed that often has a red-rusty stained upper surface.

As SE-4 and the subjacent coarse, vuggy dolostone is traced to the east of the Lake Simcoe region – it transitions into cycle V of McFarlane (1992). In this region the green marker bed was named the Pittsburgh Quarry bed by Conkin (1991). In New York State, the subjacent dolostone forms the cap of the uppermost member of the Pamelia Formation referred to as the Depauville Member by Cornell (2001). In this region the Pittsburg Quarry bed becomes substantially less well-indurated and often weathers to shaly-nodular rubble. For comparison see the facies as shown in **figure 11**. Further to the south in the Black River Valley, the green marker bed/Pittsburgh Quarry Bed can be found at the base of the Lowville Formation as far south as Boonville, New York where it appears lithologically identical to exposures in the Lake Ontario region. Cores from the Barrett Paving Quarry, and the Hawkinsville Dam on the Black River nearby, show no evidence of the lower green marker bed but show a well-developed green marker bed above fossiliferous middle Pamelia strata where it is overlain by an intraclastic dolomitic limestone often with quartz pebbles and even a few igneous rock clasts. Further to the south the unit is not present and it appears to have been truncated below the conglomeratic layer of the overlying Lowville.

Outside of the New York-Ontario region, the glauconitic, quartz-rich, dolomitic Depauville-Pittsburgh Quarry interval is chancier to recognize. Nonetheless, SE-4 approximates a position in the Camp Nelson Limestone in the subsurface of Ohio where Stith (1979) located a quartz-rich interval in the subsurface. Stith locates two marker intervals that appear very similar to the marker intervals recognized in Ontario. The lowest is referred to as marker bed II and marker bed I with MBII representing a widespread easily recognized bioturbated peloidal wackestone interval and MBI representing a two to three meter-thick argillaceous limestone and dolostone respectively. The latter is characterized by black carbonaceous papery shales and somewhat greenish dolomitic beds. In the subsurface, there is very little evidence of fossils, nor is there evidence for an associated conglomerate above the general interval. However, given the closely linked bioturbated peloidal wackestone in fairly close proximity to the argillaceous beds (i.e. MBII and MBI), this succession closely resembles the facies succession in the Ontario-New York region as discussed previously. This SE-4 is herein interpreted to be represented in Ohio at this position.

Farther south in central Kentucky, where the Camp Nelson interval is exposed at the surface a distinctive green argillaceous dolomitic interval occurs well below the Deicke K-bentonite. In the vicinity of the type section for the Camp Nelson and at Boonesborough, Kentucky the distinct green beds are easily recognized. Section on KY Rt. 27 just north of the Kentucky River and outcrop sections on Kentucky 627 at Boonesborough show a ~1 meter-thick interval that is recognized and correlated by Ettensohn et al. (2002d) and Kuhnhenn and colleagues (1981) between several sections and is referred to as the "mudstone marker." It sits just below the Oregon Formation and its characteristic "leopard skin dolostone." It is described as an argillaceous, perhaps bentonitic, limestone with "teepee" structures (Kuhnhenn et al., 1981). It is overlain by a 0.5 meter-thick leopard skin dolostone bed that has cryptalgal laminations, mudcracks, and a substantial intraclastic layer at its base. The subjacent teepee bed is interesting in that it also is associated with minor chicken-wire fabrics and evaporite mineral pseudomorphs similar to those observed in beds associated with the upper green marker bed in Ontario, which is clearly associated with a number of evaporite textures. The combined appearance of very early, if not primary depositional, dolomites in the Oregon Formation together with mudcracks suggest perhaps that these are true teepee structures produced during the formation of evaporites. Thus herein it is interpreted that this particular layer is equivalent to SE-4, and the MBI of Stith (1979).

In the upper Mississippi Valley, the position of SE-4 is intriguing and not entirely well-established particularly due to significant erosional truncation, non-deposition, and/or extreme condensation. In the upper Mississippi Valley region, contrary to the rest of the GACB, the interval within which the Deicke and Millbrig occur is almost entirely composed of dark grey to green to brown shales of the Spechts Ferry Shale. However, correlations based on the position of the Deicke K-bentonite from Kentucky assist in establishing the relative position of a dolomitic argillaceous interval at the base of the Quimby's Mill Formation. In portions of northern Illinois Basin, in northern Illinois, and southwestern Wisconsin, the Quimby's Mill is divided into three members – the basal member is referred to as the Hazel Green Member and is recognized as a massive dolostone bed that commonly weathers white (Templeton & Willman, 1952). This unit is overlain immediately by a foot of medium to dark brown shale at the contact with the

Shullsburg Member (Ludvigson et al., 2004). Within the next meter, the Shullsburg is interbedded with burrow-mottled calcareous mudstones and wackestones. The brown shales become sparse upward into the uppermost member of the Quimby's Mill, but the somewhat coarser grained interval (wackestones-packstones) still contain 5-6 cm of brown shales. Intriguingly this interval is recognized to record what Ludvigson and colleagues (2004) have named the "Quimby's Mill Carbon Isotope Excursion," which occurs beneath the level of the Deicke K-bentonite. Thus, as described, the dark carbonaceous shales sitting in the interval between the Hazel Green and the Shullsburg Members sits in approximately the same position as the Boonesborough green teepee bed, which is time equivalent with the Pittsburgh Quarry Bed and the upper Green Marker Bed of the type Black River region.

In the southern Appalachian Sevier Basin region, siliciclastic rich interval number four (SE-4) occurs near the end of flysch-style deposition and coincident with reddish limestones and molasse deposits of the Moccasin Formation and just prior to deposition of the Walker Mountain Sandstone at the base of the Bays Formation. Thus, in this region, deposition of SE-4 appears when deposition originating from the Blount Highlands was becoming subaerial. Again, the recognition of the Deicke K-bentonite (Haynes, 1992) in this region helps to establish the approximate interval within which the SE-4 would likely occur. In similar motif to the underlying SE-3 in some areas, Haynes (1992) recognizes a conglomeratic base to the Walker Mountain Sandstone that immediately overlies a shaly lime mudstone interval of the uppermost Witten Formation. Also, in southwestern Virginia to the northeast of the Tazewell Arch, Kay (1956) reports the occurrence of 1.3 meters of calcareous quartz sandstone 4 meters below the Deicke (C-3) K-bentonite. This is likely the distal equivalent of the Walker Mountain Sandstone. This sits immediately above grayish-green to reddish grey, shaly limestones of the

Moccasin Formation (Hardy Creek member) at the top of which is an intraclastic limestone conglomerate. The shales and sandstones appear to be derived from the Blount highlands to the south and east of the Tazewell Arch and the carbonates appear to be derived from areas to the west. Nonetheless, the siliciclastic-rich interval of the Moccasin Formation, capped by the conglomerate and quartz-sandstone of the Walker Mountain, help to establish this as the equivalent of SE-4. Farther south in northern Alabama, the Walker Mountain Sandstone equivalent, the Colvin Mountain Sandstone, sits immediately above the shale-rich Greensport Formation, which mimics the pattern in Virginia. Thus, it appears that it is possible to recognize the stratigraphic position of SE-4, even if SE-4 sits within an overall package rich in siliciclastics derived from the Blount Highlands during the Blountian Phase of the Taconic Orogeny.

Immediately west of the Sevier Foreland Basin, on the GACB proper, the uppermost Lebanon Limestone of the Stone's River Group is well known for its rich, diverse echinoderm faunas. Its stratigraphy has been worked out in several localities and has been correlated in the Nashville Dome region where it is divided into an upper and lower interval. The uppermost beds of the upper member were recognized by Safford (1851) to be somewhat thinner-bedded than underlying more massive and crystalline packstones and calcarenites. They are especially set off due to the presence of a number of brownish-black shales that are interbedded with the thin ribbon limestone beds in the upper 5 meters of the interval. Wahlman (1992) indicated the occurrence of dolomitic and shale-filled burrows near the top of the formation and identified that the top is disconformable with the overlying Carters Limestone Formation. The Carters is known to contain the Deicke K-bentonite (Haynes, 1992; Kolata et al., 1996) and given the unconformable contact between the Carters and the underlying Lebanon with the increased abundance of shales in the top of the Lebanon- this interval is also again in the typical position of

SE-4 and is herein considered equivalent. Clearly this region, with its relatively close proximity to the Sevier Basin, likely derived siliciclastic sediments from the Blount Highlands.

Outside of the confines of the siliciclastic influence of the Blount Highlands, the GACB region in northern Virginia and Pennsylvania are relatively pure carbonates and it is difficult to see any evidence for substantial siliciclastics. This is especially true in the central Pennsylvania region where the Benner Limestone-Linden Hall interval is extremely carbonate-rich. In fact this region produces nearly 99% pure calcium carbonate and is only occasionally contaminated by K-bentonite horizons that are typically yellowish in coloration. In a few settings, the upper Stover shows stylolitic contacts partings that enhance the appearance of some dark carbonaceous partings.

Conversely across the North Mountain hinge in the Cumberland Valley of southern Pennsylvania and northern Maryland, the Kauffman Member of the Mercersburg Formation was documented by Craig (1949) to contain an interval of four K-bentonites that appear to be equivalent to a swarm of K-bentonites in northern Virginia that includes the Deicke K-bentonite. They sit immediately above an interval of platy-bedded limestones with olive-drab to green shales and siltstones that appear to be a northern featheredge equivalent of the Walker Mountain Sandstone. Immediately below the Kaufman is the Housum Member of the Mercersburg Formation. The uppermost three meters of the Housum is very thin yellowish-gray weathering limestones interbedded with argillaceous carbonates (Craig, 1949). In the Chambersburg, Pennsylvania area, the lower Housum sits above a very widespread and well-developed intraclastic carbonate breccia with several distinctive lithologies represented (Doylesburg Member of the Shippensburg Limestone). This breccia grades upward into the Housum where massive calcilutites become distinctly bioturbated and mottled with occasional dolomitic burrow fills. Thus the distinctive transition from widespread intraclastic beds to massive bioturbated beds to platy-weathering shaly carbonates capped with siltstones mimics the pattern observed in southern Virginia and is similar to the succession in the Jessamine Dome region of Kentucky (except where the siltstones are replaced by massive "leopard -skin dolomites"). As such, it is inferred that the upper Housum platy beds is the equivalent of SE-4.

Siliciclastic Event 5

The second to last siliciclastic-rich interval, preceding the major invasion of siliciclastics at the base of the Trenton, is fairly distinctive and recognizable in the interval between the Deicke and the Millbrig K-bentonites in the Jessamine Dome region. It appears immediately above the level of the widespread faunal recurrence interval dominated by both solitary and colonial corals, including the large tabulate *Foerstephyllum*, the rugosans *Streptelasma* and *Lambeophyllum,* and the easily recognized *Tetradium*. As defined, the uppermost Tyrone Formation, below the level of the Millbrig K-bentonite is documented previously (Brett et al., 2004) to contain a 1.5 meter thick interval of green shales interbedded with minor ribbon micrites. Brett and colleagues (2004) referred to this interval as the "Marcellus" Shale (**Figure 13**) for exposures near the Garrard County line just east of Herrington Lake in west-central Kentucky. The green shales are typically grayish green to green on fairly fresh surfaces but can becomes somewhat more tan to buff in color on well-weathered outcrops. Due to its highly argillaceous composition it typically weathers back in recess relative to the pure calcilutites and coarse-grained packstones sitting immediately below and above it respectively. The Herrington Lake shale has a middle micritic limestone bed that typically has small rounded intraclasts and peloids that may have occasional oncolitic nodules. The shales themselves are typically mud-cracked, and the limestone at the base of the shales is commonly stromatolitic with 10-20 cm-wide convex upward domes that

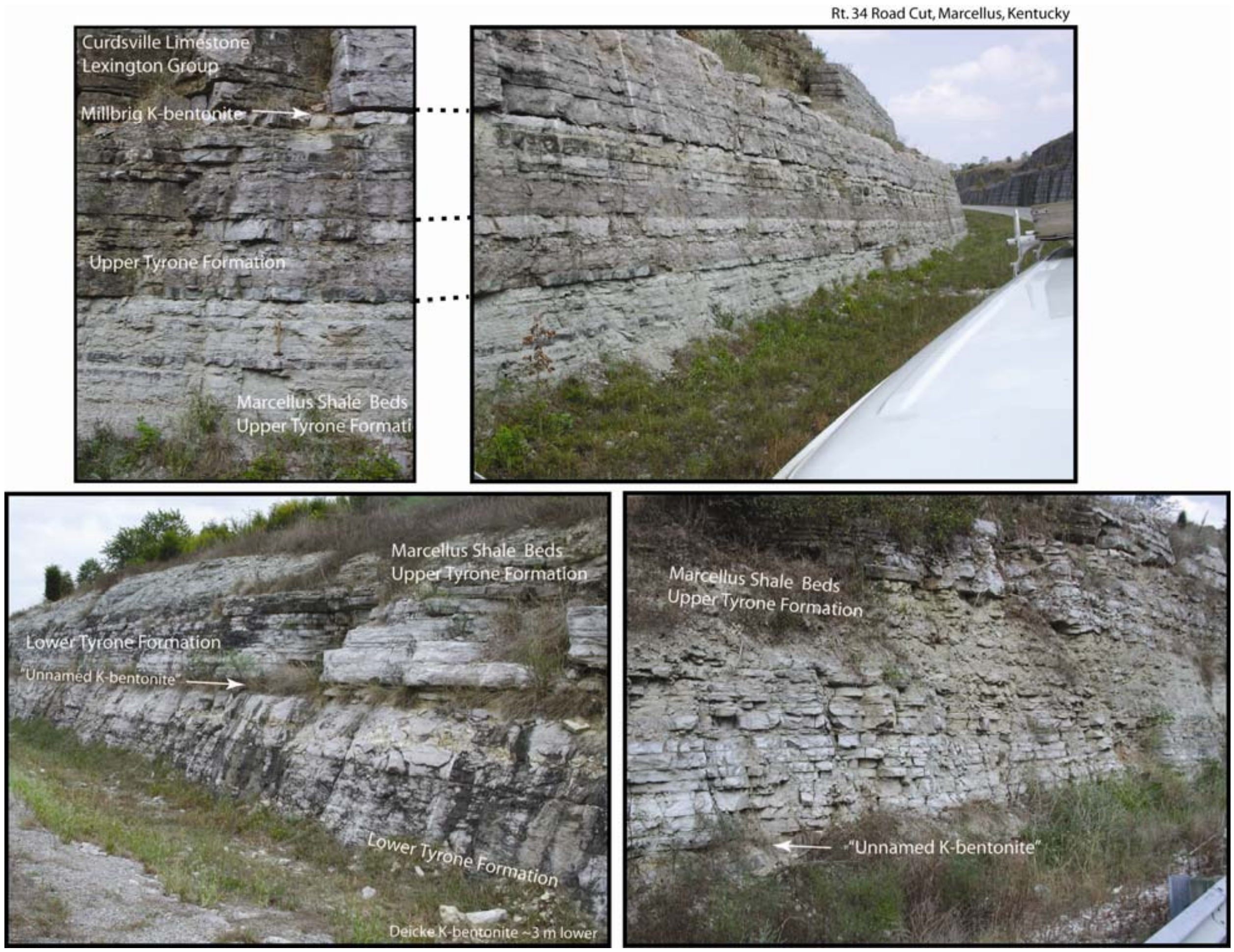


Figure 13: Road cuts in the Upper High Bridge and Lower Lexington Groups along Rte 34 in Garrard County, Kentucky. Greenish weathering shale-rich carbonate beds are herein referred to informally as the Herrington Lake Shale of the Tyrone Formation. Not the position of the Marcellus Shale relative to the Millbrig K-bentonite above and the "unnamed K-bentonite." The Deicke K-bentonite is exposed in this locality below the level of the ditch where it drops down into the Herrington Lake Gorge some 6-7 meters below the "unnamed K-bentonite."

sometimes become amalgamated and are somewhat larger. Very few of these stromatolites have grown to any significant height and therefore they are not typical domal stromatolites, but more typical of slightly hemispherical forms.

The Herrington Lake bed is not present in all localities in this region, and is missing where the Millbrig K-bentonite and upper Tyrone have been removed by erosion at the base of the Curdsville Limestone. Nonetheless, the Herrington Lake bed has been recognized on the western margin of the Jessamine Dome where it passes into the Moorman Syncline (Ettensohn & Pashin, 1992). It is also well-developed to the southeastern side of the Jessamine Dome where it is well developed at Boonesborough. In this location the Marcellus was deposited on the margin of the Rome Trough that may have been locally active during the Upper Ordovician. In this region the Millbrig has not been recognized and appears to have been truncated, but erosion appears not to have cut down to the level of the Marcellus. It is also known from the Tyrone type-region, but is not well developed in the interval at Camp Nelson that straddles the Cincinnati Arch proper. In this locality on KY Rte. 27, the Millbrig has been identified albeit in cryptic and sometimes laterally discontinuous pockets. In this locality, the Marcellus is likely equivalent to a thin interval of argillaceous, mud-cracked platy limestones, rich in *Bathyurus* trilobites and low, relatively small convex stromatolites.

Stith, (1979) recognized several markers in cores from southern Ohio that could be correlated on the basis of wireline logs. In most cases, however, his study does not recognize either the K-bentonite that sits between the Deicke and the Millbrig nor does it recognize the shaly Marcellus interval. Cores observed from the northern Cincinnati region and from Butler County, which overlaps with the region investigated by Stith, contain both the "unnamed" k-bentonite between the Deicke and the Millbrig, and a zone of thin ribbon micrites with thin green shale stringers reminiscent of the Marcellus beds (Cornell et al., 2001). The Marcellus interval is also recognized in the cores taken in the Mill Creek by the Army Corps of Engineers. It is again composed of the typical greenish ribbon micrites and interbedded mud-cracked shales all below the level of the Millbrig.

In the upper Mississippi Valley region, the occurrence of SE-5 as its own event is much more difficult to establish owing to the substantial Decorah Shale in north eastern Iowa – and southern Minnesota. This unit is relatively thin to almost absent on the Wisconsin Arch, but expands to the northwest in Minnesota and westward in the subsurface of Iowa into a much more substantial thickness of shales that encompass the entire upper Black River through Middle Dunleith Formation (Ludvigson, et al., 2004; 2005). However, it is possible to establish the position of SE-5 using the well-correlated and fingerprinted K-bentonites including the Deicke and Millbrig (Emerson et al, 2004). Whereas in central Kentucky the Deicke and Millbrig are separated by nearly 10 meters of relatively pure shallow-water carbonates, on the Wisconsin Arch in the vicinity of the Guttenberg type-section and at Dickeyville (in Wisconsin), the Deicke and Millbrig are separated by less than 0.4 meters of shaly nodular carbonate (Carimona Member of the Spechts Ferry Shale). Further north in the vicinity of Rochester, Minnesota, their separation increases up to 4 meters with substantially more shale separating the two. In this region, the Carimona is about 1.5 meters thick and is overlain by dark shales with a few interspersed shell beds of the Glencoe Member of the Spechts Ferry Shale (Emerson et al., 2004). Again this interval is condensed and truncated on portions of the Wisconsin Arch; however, shales dominate deposition in the Hollandale Embayment to the northwest. Given the evidence for extreme condensation, it is difficult to establish the exact timing of shale deposition on the Arch, but it appears that siliciclastic dominated deposition commenced following deposition of the Platteville Limestones (Quimby's Mill Formation) sometime after deposition of the Deicke K-bentonite but prior to deposition of the Millbrig, Certainly north and westward of the Wisconsin Arch, shale deposition was ongoing for some time before the Millbrig was deposited as at least 2 meters of basal Glencoe Shales separates the nodular Carimona from the Millbrig (Emerson et al., 2004). Thereafter, especially in Minnesota, western Iowa and even eastern South Dakota – shale deposition continued to dominate facies nearly continuously well up into the Dunleith (Witzke and Ludvigson, 2005).

Given the dominance of shales in the former region, recognizing a time equivalent shale deposit to the siliciclastic-rich intervals found in the vicinity of the Cincinnati Arch and elsewhere is a challenge. Nonetheless, Ludvigson and colleagues (2004) report an interval of distinctively greenish shales below the level of the Millbrig and above the Deicke in eastern Iowa. These stand out as most of the shales in the stratigraphic succession are decidedly brown to dark brown in color. The appearance of greenish shales in this specific stratigraphic position approximates the position of the Marcellus interval in central Kentucky. Also distinctive is that Ludvigson and colleagues (2004) report the simultaneous occurrence of a positive carbon isotopic excursion below the Millbrig within this greenish interval as detected in a core from Clayton County, Iowa. These authors refer to this excursion as the Spechts Ferry Isotopic Carbon Excursion and the shift precedes the Guttenberg Isotopic Carbon Excursion. Thus it appears that within the overall succession of Decorah Shales, there is a potential candidate for SE-5 as constrained by K-bentonite occurrences.

In southern Iowa, and western Illinois, the Deicke and Millbrig are missing along the flank of the Wisconsin Arch/western edge of the Illinois Basin, but both return in successions to the south of Iowa in northeastern Missouri and southwestern Illinois. In this region, a core from Sangamon County, Illinois places the Deicke at the contact of the Quimby's Mill/Macy Limestone below and the Castlewood Member of the Spechts Ferry above (Ludvigson et al., 2004). The Spechts Ferry is in turn overlain by the coarse-grained Kimmswick Limestone. Thus, the Castlewood is an equivalent of the Carimona further north. In the western Illinois, the upper Castlewood is dominated by massive, burrow-mottled, fine-grained carbonates, but the upper portion becomes wavy nodular with minor shales and is capped by ~30 cm of phosphatic shale. This in turn is sharply overlain by a coarse-grained fossiliferous limestone facies (similar to the Curdsville Limestone of Kentucky) of the lower Kimmswick. No Millbrig is present below the sharp contact with the grainstone facies of the Moredock Member of the Kimmswick Limestone and is likely truncated. Evidence for SE-5 is not indicated in the data presented by Ludvigson and colleagues (2004); however, they do recognize the Spechts Ferry Excursion among the wavy nodular upper Castlewood.

In the Ontario-New York Region, successions in the upper Black River Group also contain evidence for SE-5. In this region, as in the Tyrone of Kentucky and the upper Castlewood of Missouri, facies of the Moore Hill/House Creek Limestones (uppermost Gull River; Upper Lowville respectively) are characterized by well-developed, massive, burrow-mottled micritic mudstones. This interval is also known for very large *Foerstephyllum* corals that are a part of the recurrent fauna discussed earlier and is representative of significantly deeper deposition (but still shallow sub-tidal conditions compared to *Tetradium*-rich facies). The top of the Moore Hill/House Creek interval contains an easily recognizable shaly limestone interval named the Weaver Road Beds (Cornell, 2001; Brett et al., 2004). The Weaver Road Beds are distinctive in that they are typically thin wavy-bedded ribbon limestones with thin brown to green laminated shales (**figure 14**). In well-developed areas, this unit consists of a tripartite succession with a base of ribbon micrite beds succeeded by stromatolitic beds. The latter are often very large (up to 1 meter high) domal stromatolites and may even become thrombolitic in some cases – including in the Chaumont Quarry. The uppermost part returns to ribbon micrite beds that are interbedded with greenish shales that can weather to yellow tan or to brownish shales that contain substantially more organic matter. Most commonly this interval is mud- cracked and micrite interbeds may have imbricate intraclasts and on some bedding planes there are often entire specimens of *Bathyurus extans* trilobites. The Weaver Road beds show evidence of

Figure 14: Exposures of the Weaver Road beds in Jefferson County, New York. As shown are photographs from 6 different localities where the Weaver Road Beds are exposed.

truncation in some localities such that the upper ribbon micrites and the medial stromatolitic beds may not be present beneath the base of the Leray, which is a distinctly coarser and darker gray grainstone bed – that in some localities contains small disseminated quartz-sand grains.

In Ontario, the Weaver Road equivalent beds are well developed (**figure 15**). In the region north of Napanee where Kay (1931) designated the type section for the Glenburnie Shale. The Weaver Road bed interval occurs below the Glenburnie and the Leray. In the Kingston region, the Weaver Road interval exhibits typical tripartite components with the central domal stromatolite interval. Further west, in the Napanee region, the interval is dominated by ribbon micrites on the southern margin of Napanee where the unit is exposed on Hwy 2. Just off the

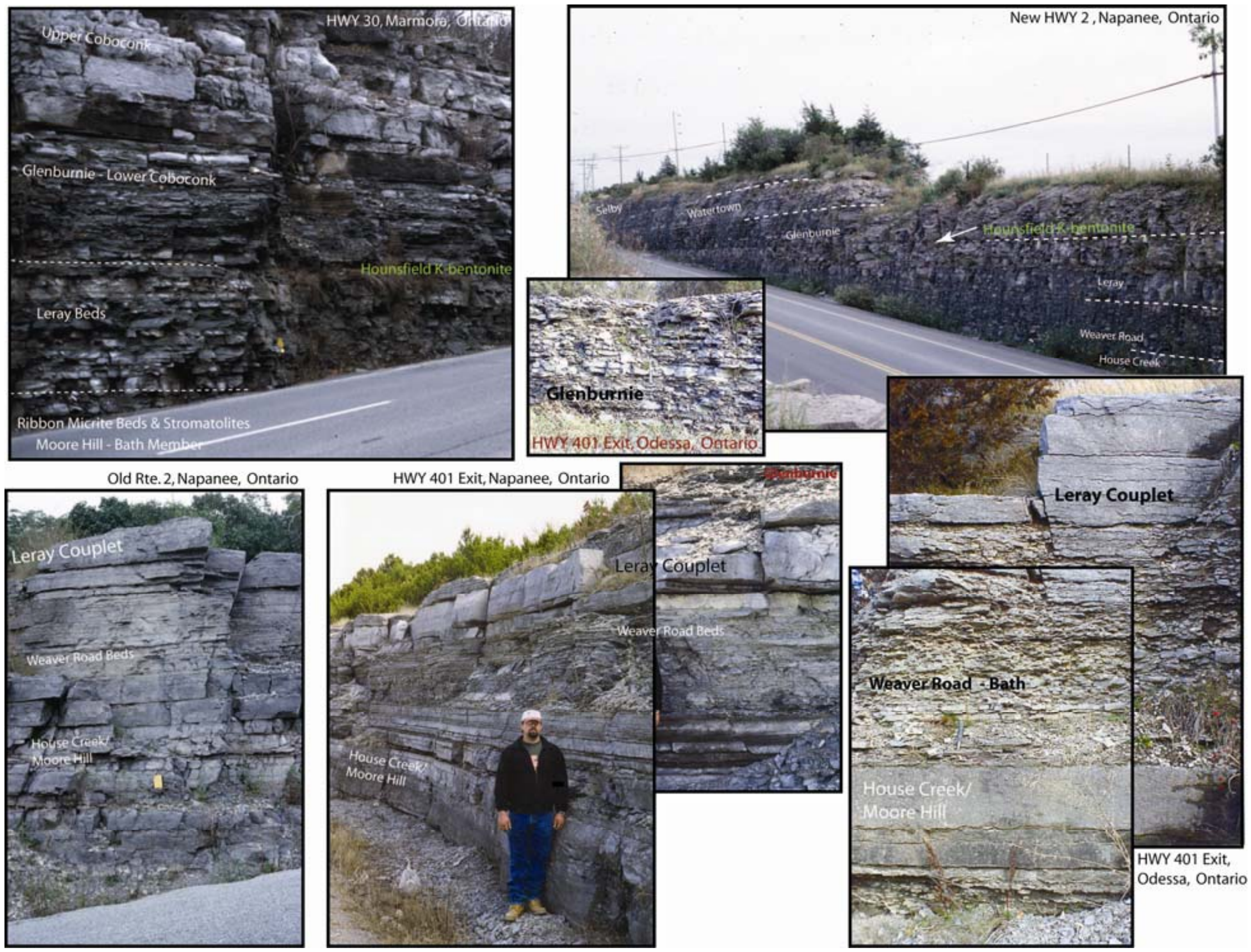


Figure 15: Outcrop exposures of the Weaver Road-Leray-Glenburnie-Watertown succession as exposed in Ontario ranging from the Napanee, Ontario region westward to Odessa and Marmora, Ontario.

exit ramp from Hwy 401, the same interval is dominated by platy-weathering greenish-gray shales. In both instances there is only minor evidence for microbial laminations, which occur in some of the smaller ribbon beds. Slightly domal stromatolites are also found at the base of the platy shales and are more reminiscent of the forms found in the same interval in Kentucky. Aside from cyanobacterial mats, the Weaver Road equivalent (Bath Member of Noor, 1989) in this area is typically barren except for some bedding planes with numerous ostracods, small ctenodont bivalves, and occasional fragments of *Bathyurus* trilobites. As in New York, mudcracks are also common. Thus in the New York – Ontario region, it appears that the Weaver Road interval is coincident with the fifth siliciclastic event (SE-5) that coincides with a significant shallowing at the end of the Lowville back to peritidal and even supratidal conditions prior to a pronounced facies dislocation into deeper-water facies of the Leray.

In northeastern New York State in the Lake Champlain region, Bechtel (1993) recognized the House Creek interval at the top of the Lowville Formation and recognized the same characteristic facies and faunas identified by Walker (1973) in the type region. Immediately overlying the House Creek and immediately below the Isle La Motte is a unit named the Sawyer Point Formation. The unit stands in sharp contrast to the light-to medium grey of the House Creek below. On South Hero Island in northern Lake Champlain, the Sawyer Bay is characteristically lumpy to nodular-bedded, with wavy to slightly crinkly lamination and occasional ripple cross-lamination. It has a brachiopod-trilobite dominated biofacies composed almost entirely of very dark gray to black micrite with thin biosparite lenses and occasional layers of lenticular chert nodules. This unit is interpreted by these authors to represent localized and very rapid deepening of the Lake Champlain area immediately following karstification and exposure of the upper House Creek.

The Sawyer Bay thus sits approximately in the same position as the Leray of western New York and in terms of coloration and lithologic similarities, resembles the latter unit. To date, a Weaver Road equivalent interval below the Sawyer Bay has not been recognized, and its former location may be represented in the karstic unconformity at the base of the Sawyer Bay. Considering petrographic evidence presented by Bechtel (1993) that show multiple episodes of early marine and meteoric cements, the inference is made that major extensional block faulting initiated in the Lake Champlain region at the time of the unconformity and during deposition of the Sawyer Bay. With localized uplift and subsidence, initiating during the deposition of the upper House Creek, these uplifts along the Adirondack Arch region may have been enough of a structural feature to produce the restricted and protected conditions under which the Weaver Road interval was deposited in more westerly locations.

As in northeastern New York State, the approximate position of the Marcellus-Weaver Road interval has not been expressly identified in outcrops in central or southern Pennsylvania. Wagner (1966) using cores and wireline logs recognized a shaly-interbedded limestone interval at the base of the Nealmont Formation of the Bedford County region in south-central Pennsylvania near the West Virginia border on the southeastern margin of the Rome or Olin Troughs. The argillaceous – calcisiltite interval occurs below three bentonites recognized in the Centre Hall member of the Nealmont. He uses the higher gamma ray values in tandem with well cuttings to distinguish the contact between the Benner Formation and the Nealmont. Wagner does not report the Curtin (Upper Linden Hall) Limestones (Valley View, Valentine, Oak Hall) in his discussion and as indicated by Kay (1944) the unit is often truncated below the Nealmont Formation. Therefore, the stratigraphic position of the argillaceous-calcisiltite interval is not well-constrained – but it may in some part represent a combined condensed equivalent of SE-4 (Pittsburgh Quarry Bed) or it may represent SE-5 and be equivalent to the Weaver Road-Marcellus shale-rich interval. However, the detailed well descriptions from the Bedford County region show a 9.5 meter thick interval at the top of what Wagner (1966) refers to as Benner Formation. The interval is composed of cryptocrystalline limestone with birdseye structures, minor shell hash layers and traces of K-bentonites in at least two different intervals. The lithologies are identical and representative of the Valley View (6.5 m)-Valentine (~3.0 m) of the Nittany Valley northwest of the Nittany Arch and very similar to facies in the Lowville of New York. Moreover, these lithologies immediately underlie the argillaceous zone at the base of the Nealmont. Thus it is tentatively suggested that the argillaceous-calcisiltite interval sits in a position that is closely allied with the upper Lowville and may represent the approximate position of the Weaver Road-Marcellus shale.

In the Cumberland Valley of southern Pennsylvania, Craig (1949) recognized the Kauffman Member of the upper Mercersburg Formation as a dark-gray, crinkly to slabby, weathering limestone. In outcrop exposures they are commonly fossiliferous wackestones and packstones with occasional *Tetradium* and other fossil fragments. The interval is commonly well-bioturbated and can become substantially more massive in some localities. The massive beds in the lower to middle part of the formation contain four distinct K-bentonites Near the top of the formation, the beds are thinner and in more westerly outcrop exposures weather in thin platy-sheets that weather to a buff yellow color (Craig, 1949). On fresh surfaces the shaly partings separating the crinkly-laminated beds are black, papery, carbon-rich partings that contain no discernable fossil fragments. Overlying beds have been referred to as the Greencastle Member of the Chambersburg Formation. The latter unit is yet another massive bioturbated, dark gray to black limestone with occasional K-bentonite partings. The 20 meter thick unit shows an abruptly deepening-upward succession with a thin calcarenite or grainstone bed at its base that is locally developed into an intraclastic, flat-pebble conglomerate facies. It is a lateral equivalent of the Nealmont Formation and becomes distinctly shalier at the top where it is equivalent of the Oranda Formation of northern Virginia region. Thus the uppermost beds of the Kauffman, below the level of the Greencastle, although lacking significant argillaceous components, are likely equivalent to beds attributed to the SE-5 interval.

As in the upper Mississippi Valley, SE-5 of the central carbonate platform appears to correlate into an expanded interval of siliciclastic-influenced carbonates in the Eggleston Formation of southwestern Virginia. Kay (1956) described a siliciclastic to mixed carbonate interval in the Edinburg Formation of this region. Located above Rosenkrans' (1934) V-3 K-bentonite (Deicke equivalent per Haynes, 1992; Kolata et al., 1996) and below V-4 (Millbrig per Haynes, 1992; Kolata et al., 1996), Kay (1956) identified an 11 meter-thick argillaceous calcilutite interval floored by a 0.6 m thick quartz sandstone capped by thin-bedded, crinkly-laminated micrites interbedded with thin shales (ribbon-limestone facies). The ribbon beds commonly weather buff-greenish yellow and are capped by a 30 cm thick dark gray to black silicified limestone with rippled upper surface. This bed, in turn, forms the floor of the 30 cm thick V-4/Millbrig K-bentonite. Given the recognition of the V-4 as the Millbrig and the V-3 as the Deicke; the position of the argillaceous ribbon-limestone interval is confidently placed in an equivalent position to the "Herrington Lake shales" of Kentucky. In this region, lower to middle Edinburgh strata, although dominated by relatively pure limestones, show an appreciable siliciclastic influence (quartz sandstones, siltstones, and shales) throughout the interval and reflect deposition during waning phases of the Blountian tectophase immediately prior to activation of the northern or Vermontian tectophase.

Siliciclastic-rich interval five is also recognized in the Carter's Limestone of central Tennessee. Following on previous authors, Wahlman (1992) divided the Carter's into a lower and upper member with the boundary roughly established at the position of the Deicke K-bentonite (B-3 K-bentonite). The upper member is thus recognized and defined as a dense, fine grained, laminated, argillaceous limestone extending up to the base of the Hermitage Formation with the contact located immediately above the position of the Millbrig K-bentonite (B-6 K-bentonite; Wahlman, 1992). In outcrops in the central to northeastern Nashville Dome (Gladeville, Carthage, and points to the northeast), the upper member of the Carters contains a thin 1-3 meter thick interval of green mudstones to burrowed lime mudstone facies interbedded with laminated dolomitic limestones as described by Holland & Patzkowsky (1998). The green shaly facies is well developed at Gladeville (where the Millbrig is absent by unconformity), but is also present at Bluff Creek where the bed is found below the Millbrig. As described SE-5, in the Nashville Dome region is comparable to facies found in the High Bridge Group of central Kentucky.

Siliciclastic Event 6

The final pre-Trenton siliciclastic-rich interval (SE #6) is designated on the basis of facies in the upper Mississippi Valley region in the vicinity of the Millbrig K-bentonite. As mentioned in discussion of SE-5 for the upper-Mississippi Valley, this region was impacted by a fairly long period of time in which deposition was dominated by siliciclastics –initiated after deposition of the Deicke K-bentonite and continuing until deposition of the Garnavillo Member of the Guttenberg was initiated (Ludvigson et al., 2004). In eastern Iowa, the Millbrig sits within the Glencoe sub-member of the Spechts Ferry Member of the Decorah Formation (Witzke and Glenister, 1987). The interval immediately above the Millbrig is characteristically condensed and shaly in Iowa and is typically medium to dark brown shales that become interbedded with nodular packstone beds on the flanks of the Wisconsin Arch in the vicinity of Dickeyville (Emerson et al., 2004). These shales typically have stringers of brachiopods and bryozoans, and often contain burrow networks.

The Glencoe is capped by the distinctly more fossiliferous and massive mudstone-wackestone carbonates of the Garnavillo Member of the Guttenberg Limestone (Ludvigson et al., 2004). The Garnavillo is distinctive in that it often contains a 20-30 cm basal wackestone to packstone bed with phosphate nodules immediately below the position of the Elkport K-bentonite. In many places including in Illinois and southern Wisconsin, the upper Glencoe shaly-interval is truncated below the Garnavillo phosphate horizon and the contact was interpreted to represent a drowning unconformity (Kolata et al., 2001; Ludvigson, 2004). To the northwest into Minnesota and into western Iowa, the supra-Millbrig portion of the Glencoe expands into a substantial thickness of shale referred to as the Decorah Shale – although the Fe-rich, phosphatic Garnavillo bed below the Elkport can apparently be traced out into the shales of this region (Ludvigson et al., 2004; Witzke and Ludvigson 2005). The same Glencoe shale - phosphatic bed interval can also be traced to the south and east of Iowa where it is seen as the basal bed of the Kings Lake member of the Kimmswick Limestone, even where the Millbrig and Deicke are not observed. In this case, SE-6, the upper Glencoe appears to be a fairly widespread unit in this region. Although siliciclastic deposition continued to the north – the siliciclastic event designated as SE-6 is thus confined to the interval immediately surrounding the Millbrig K-bentonite and extending up to the base of the Garnavillo Member of the Guttenburg Limestone and ranges from a few 10's of cm on the Wisconsin Arch to 2 meters or more in eastern Iowa and southern to northern Minnesota (Witzke and Ludvigson, 2005).

Outside of the upper Mississippi Valley region, the interval immediately above the Millbrig K-bentonite is often dominated by relatively pure to coarse grained packstones to grainstone lithologies. In some cases, these facies are known to overstep underlying beds including the Millbrig K-bentonite (Brett et al., 2004). The interval is dominated by coarse-grained facies in the Jessamine Dome area of Kentucky, where the entire interval is referred to as the Curdsville Limestone Member of the Lexington Formation. Although there are silty shales near the basal contact in the vicinity of the Millbrig they have been considered to be relatively minor (Cressman, 1973). Nevertheless, given the importance of siliciclastics in the vicinity of the Millbrig in the upper Mississippi Valley, the interval surrounding the Millbrig in Kentucky has been re-studied. **Figure 16** shows two intervals of the basal Curdsville Limestone from two localities in the Jessamine Dome. Close inspection of the succession at Marcellus, Kentucky reveals a green silty-argillaceous calcilutite facies in the 60 cm interval below the Millbrig, and approximately 40 cm above the Millbrig another thin 20 cm interval of silty shales and calcareous laminated siltstones is recognized below a widespread intraclastic conglomerate bed. The same interval is recognized at Shakertown, Kentucky where there is evidence of erosion and channeling in the basal Curdsville such that the silty shale and laminated siltstone interval are truncated in some locations along the contact immediately below the intraclastic grainstone horizon. Also significant is the appearance of the phosphatic, iron stained cap at the top of the intraclastic bed. This 1m thick interval is approximately equivalent to the upper Glencoe and Garnavillo Member of the Guttenburg from the Upper Mississippi Valley.

In the type Black River region, minimal evidence exists for pre-Millbrig shales. However the Leray Formation in a few localities, especially in well-weathered and etched outcrops in the Black River Gorge, demonstrates a discontinuous sandy wackestone that commonly weathers buff tan to orange with an iron-stained upper surface. This, when present, stands in contrast to the white and dark grey beds immediately above and below. The overlying burrowed-nodular wackestone contains minor disseminated coarse-silt to fine-sand quartz grains. This bed sits immediately below the Hounsfield K-bentonite (sensu Kay, 1931; and the probable equivalent of the Millbrig, but see Mitchell et al., 2004). Immediately overlying the Hounsfield K-bentonite is a pack- to grainstone bed approximately 50 cm thick that is capped by silty shale partings that grade upward into a burrowed-nodular wackestone to grainstone with variable amounts of clay.

In most Black River Gorge localities, especially in the vicinity of Watertown and upstream in the

Rte. 34 Outcrop, Marcellus, KY
intraclastic horizon
silty-shale & laminated siltstones
Curdsville Formation
packstone-grainstone
Millbrig K-b
birdseye micrite
silty-argillaceous calcilutite
Tyrone Formation
Rte. 68 Outcrop, Shakertown, KY
rudstone-grainstone
phosphatic-fe-stained cap
intraclastic horizon
laminated siltstones
Curdsville Formation
packstone-grainstone
Millbrig K-b
Rte. 34 Outcrop, Marcellus, KY
Capitol K-b
Curdsville Formation
rudstone-grainstone
phosphatic-fe-stained cap
planar laminated grainstone
intraclastic horizon
silty-shale & laminated siltstone partings
packstone-grainstone
Millbrig K-b
birdseye micrite
silty-argillaceous calcilutite
Tyrone Formation
birdseye micrite
Rte. 54 Outcrop, Brownville, NY
Massive wackestones to packstones
weak Fe-stained cap
grainstone
intraclastic horizon
Watertown
Thin nodular wackestones
wavy grainstones
Glenburnie
silty-shale partings
packstone-grainstone
Hounsfield K-b
Leray Member
burrow-nodular wackestone
dissseminated quartz grains
birdseye -Tetradium micrite
Black River Gorge Glen Park,, NY
Hounsfield K-b (sensu Kay 1935)
Selby
Unnamed K-b
Massive wackestone to packstone
Watertown
intraclastic horizon
Thin nodular wackestones
wavy grainstones
Intermittent Hounsfield K-b
packstone-grainstone
Glenburnie
burrow-nodular wackestone
disseminated quartz grains
Leray
birdseye -Tetradium micrite
Black River Gorge Watertown,, NY
Watertown
intraclastic horizon
Glenburnie
Shaly nodular wackestones
wavy grainstones
silty-shale partings
packstone-grainstone
Hounsfield K-b
Leray
burrow-nodular wackestone
dissseminated quartz grains
birdseye -Tetradium micrite
Weaver Road Beds

Figure 16: Outcrop exposures in the vicinity of SE #6 from Kentucky and New York showing the approximate position of the Millbrig and Hounsfield K-bentonites. Recently the Hounsfield has been shown to be the Millbrig K-bentonite (Mitchell, et al., 2004). Despite nomenclatural disagreements, the Hounsfield type locality has now been confidently correlated into a newer reference section on Rte. 54 immediately north of the type quarry. It has also been correlated into the Black River Gorge sections where exposures in the hydro-electric power generation diversion channel (when drained) show at least 2 additional K-bentonites above the Glenburnie- Watertown interval.

villages of Great Bend and Black River, acid etching helps this interval weather to a shaly nodular fabric. It becomes extremely distinctive compared to the more massive-bedding of the overlying Watertown Formation and often can become confused with the Selby Formation in limited outcrop exposures. In fresh outcrops as exposed in the water diversion channel (when drained) at Glen Park, and in quarries, the shaly nodular fabric is less distinctive. However, where well weathered, this interval is referred to as the Glenburnie Shale, and it contains facies variations not unlike the upper Spechts Ferry Shale of the Upper Mississippi Valley. Moreover the presence of an intraclastic grainstone bed immediately above the Glenburnie (often bordering on brecciated) is capped by a faint, but recognizable undulatory iron-stained surface that has some pyritic nodules and vertical borings. Thus far phosphatic nodules have not been recognized at this surface. Thus, although the shales characteristic of this interval in the Upper Mississippi Valley are significantly less prominent in New York, a number of siliciclastic-rich carbonates are developed in the vicinity of the Hounsfield. In addition, a similar motif is developed immediately above the Glenburnie interval with the mineralized surface that may represent the same surface as recorded at the top of the Garnavillo of the basal Guttenberg Formation.

In eastern New York State, the equivalent unit to the Watertown Formation is the Isle La Motte Limestone. To date, no K-bentonites have been recognized in the Isle La Motte although some potential re-entrants have been detected. In addition, due to structural complexities, local faulting, and extensive quarry operations in the Plattsburgh region, most exposures of the Isle La

Motte have either been completely removed for quarry stone or are so overgrown that their study is difficult (Fisher, 1968; Bechtel, 1993).

In central Pennsylvania, recognition of SE-6 on the basis of the occurrence of the Millbrig K-bentonite, as currently recognized (as per McVey, 1993), is problematic and contentious. Based on foregoing observations, outlined elsewhere, and the recognition of several relatively thick K-bentonites in the Nealmont Formation in several localities, the position of SE-6 is tentatively established within that unit. Kay (1944) provided detailed stratigraphic descriptions for the three members of the Nealmont which include the basal Oak Hall Member (now included in the underlying Linden Hall Formation by Rones, 1968), the middle Centre Hall Member and the uppermost Rodman Member. In the Oak Hall Quarry and in the Pleasant Gap Quarry, the interval is still very well exposed and the occurrence of N1, N2, and N3 K-bentonites have been noted and used by quarry workers. Kay (1944) reported a thickness of over 12 centimeters for the N1 and 20 cm for the N2 at Tusseyville – just a few miles east of Oak Hall. N1 sits within the Oak Hall Member and the N2 occurs at the very top. Thus, based on the similarities with the succession in New York-Ontario, the succession of litho and bio-facies in the Oak Hall to Centre Hall, combined with the occurrence of the relatively thick N2 K-bentonite, herein the basal Centre Hall shaly zone, is correlated with the upper Glencoe-Glenburnie siliciclastic event (SE-6).

In the Cumberland Valley of southern Pennsylvania, the K-bentonites of the Nealmont and the Salona have been the basis for long-standing well documented studies by Thompson, (1963). Kolata and colleagues (1996) have used Thompson's correlations to establish the equivalents of the Salona K-bentonites in outcrop sections at Harrisburg, Pennsylvania in a unit they describe as the Oranda Formation. The Oranda is not formally applied to these rock units but originates in the Shenandoah Valley in northern Virginia. Thus in the Cumberland Valley region, the interval in question is commonly attributed to the Myerstown member of the Jacksonburg Limestone. In the Lebanon and Lehigh Valley region, the Myerstown unconformably overlies St. Paul Group and even underlying units of Beekmantown age with a significant unconformity. To the south of Harrisburg the amount of time represented by the basal Myerstown Unconformity decreases through Cumberland County and into Franklin County. In the vicinity of Chambersburg, the Myerstown equivalent transitions into the upper Chambersburg Group where it was referred to as the Greencastle bed of the Chambersburg Limestone (Bassler, 1919). The lower portion of this bed appears to be relatively restricted in its distribution and is not known from western valleys where it has been truncated under the upper Greencastle – Myerstown.

Although K-bentonites have not been recognized in the Greencastle Bed the interval is characteristically shaly at its base. In the Chambersburg type locality, the Oranda equivalent bed is a 3 meter thick massive dark gray calcilutite with a few crinoid grainstone stringers and occasionally *Echinosphaerites*. The bed is underlain by nodular or "cobbly" weathering limestones, interbedded with yellowish shales (Bassler, 1919). In the type locality at Greencastle, the massive, relatively pure limestone of the upper Greencastle Member of the Chambersburg is underlain by a 2.5 m thick interval of argillaceous nodular wackestone that is sandy at the base. Below the sandy beds, the uppermost Mercersburg Limestones appear and are similar in lithologies to the Linden Hall Formation as previously mentioned. Recent reconnaissance work has suggested that some of the yellowish shales may indeed be bentonitic but their recognition is hindered by poor exposures.

Nonetheless, farther south into Maryland and close to the West Virginia border, this same interval has been shown to become substantially more bioturbated and limestone-rich and loses its siliciclastic facies. On the flanks of the Tazewell Arch, the Millbrig K-bentonite is recognized in limestone dominated facies (upper Eggleston) with very little evidence of siliciclastic influence. However, farther east the Millbrig is found within interbedded shales, siltstones, and sandstones of the Bays Formation (Haynes, 1992). In the latter region the shales have been noted to become greenish-grey and exhibit intervals of mudcracks (Kay, 1954) suggesting very shallow-water deposition under reducing conditions. This green siltstone interval of the Bays was called the “green Bays” by Haynes (1992) and contains the fingerprinted Millbrig and may represent deposition in a restricted estuarine to deltaic system originating in the well-weathered Blount Highlands to the southeast. The green-colored interval of the Bays Formation terminates just above the level of the V-7 K-bentonite and is succeeded by the massive pure carbonates of the Oranda Formation (basal Martinsburg). Thus, in this region, as in the Cincinnati Arch region SE-6 is characterized by greenish siltstones and sandstones. In contrast to siliciclastics in the upper Mississippi Valley region who are shown to be derived from the craton, it is likely that siliciclastics found in the vicinity of the Millbrig K-bentonite on the flanks of the Cincinnati Arch, and points eastward are derived from source regions off the Virginia Promontory in the Blount Highlands.

Subsequent to deposition of SE-6, carbonate production resumed across the GACB, however for a very limited time. The ensuing last stand of the GACB, i.e. the widest and most uniformly developed facies of the GACB, is uncharacteristic of most of the preceding peritidal succession where carbonate deposition dominated during transgressive and early highstand phases only to be mixed or influenced by siliciclastics in regressive phases. In the ensuing earliest Chatfieldian not only is there evidence for a major synchronous isotopic carbon excursion, but biofacies uniformly show a return to brachiopod-echinoderm-dominated faunas. Although wackestone to grainstone facies variability does occur and although there are some rapid lateral facies changes in the vicinity of tectonic induced structures, the entire succession uniformly shows a rapidly deepening-upward succession.

The deepening phase is capped by widespread deposition of allodapic carbonates interbedded with the first pulse of siliciclastics from the mafic-dominated Vermontian Tectophase of the Taconic Orogeny. Siliciclastics deposited at this time in the Upper Mississippi Valley have been shown based on neodymium isotopes (Fanton et al.; 2002) to represent the major change to orogen-derived sediments. Clearly the transport of sedimentary materials from the orogen out into the GACB across distances of up to 1300 km or more, required a substantial change in the physical architecture of the GACB and adjacent foreland basin, as well as a major change in oceanographic circulation.

Summary & Implications of Siliciclastic Events in CBRT:

In the pre-Trenton carbonate platform succession, it appears that many of these shallow-water siliciclastic events coincide with periods of significant restriction and shallowing of the GACB. Many of these siliciclastic events coincide with periods when the carbonate factory was not operating normally and likely was influenced by local topography and variations in sea- and sub-sea water chemistry and nutrient availability. This observation is problematic in that the introduction of siliciclastic sediments across widespread portions of the GACB is not likely explained by marine transport mechanisms under extremely shallow platform conditions. This is especially exacerbated in areas that are especially distant to obvious sediment source areas (i.e.

the Ouachita Uplift, the Transcontinental Arch, the Canadian Shield, or even the Blountian Highlands).

Thus, in most cases, the delivery of argillaceous materials and associated quartz sands has to be explained by alternative transport mechanisms. As is well-established the CBRT interval has become well-known for its large number of K-bentonite horizons and presumably these wind-blown volcanic ashes could have contributed a significant volume of siliciclastic sediment over wide areas. Given enough time and adequate conditions, some of these tephras could have been reworked, weathered, transported locally, and intermixed with carbonates and thus their appearance as K-bentonites becomes difficult. Nonetheless it is clear that in many cases, volcanic ash beds were preserved even within the intervals that these siliciclastic events are developed. Moreover, their occurrence near the top of shallowing-upward packages suggests an alternative sediment source is plausible.

As mentioned in discussion of the chert-rich intervals, especially the CR-3, there is significant evidence that the GACB region may have been subjected to at least intermittent periods of aridity and eolian transport of sediment. As in this last scenario, it is at least possible that the siliciclastic-rich intervals in the CBR interval are coincident with the introduction of wind-blown dust derived from the sub-tropical desert belts that likely existed in portions of the Laurentian continent and adjacent and approaching Taconic Highlands located offshore to the southeast.

Stromatolite & Ooid-Rich Intervals: Calcification Events of the CBRT

Introduction & General Background:

In addition to the chert and siliciclastic-rich intervals already discussed, there are a number of other distinctive lithologic intervals that are important to mention as they appear to have some chronostratigraphic importance. Throughout most of the Chazy and Black River (CBR) interval on the GACB, stromatolites and ooid carbonate facies have been reported in most outcrop regions in the study region. These facies are characteristically absent in the overlying Trenton Group and its equivalents, with the exception of the Hull in the Ottawa region (Kiernan & Dix, 2000). Their absence has been used as evidence supporting the idea that during deposition of the Trenton, even peritidal regions were influenced by cool-water of moderate salinity (i.e. Pope & Read, 1997a; Brookfield, 1988, Holland & Patzkowsky, 1996). Even in the CBR interval, most facies are dominated by peloidal micrites, massive wackestones, or biosparite/calcarenite facies. Thick repetitive cycles of ooid-rich facies, and/or stromatolitic boundstones are not a primary characteristic of the early Late Ordovician as compared to former Cambro-Early Ordovician carbonate settings of the older GACB. This observation, among others, was used by Brookfield (1988) to suggest that deposition even during the Black River Group may have been related to cool-water as well.

Nonetheless, ooid-containing facies, as well as stromatolite occurrences have been mentioned occasionally as accessory elements to broader facies associations from Chazy through Black River strata in the Ontario-New York region (Textoris, 1968; i.e. Walker, 1973; McFarlane, 1992; Grimwood et al., 1998, 1999; Kiernan, 1999). In most instances they have been reported to represent fairly localized facies development in the vicinity of minor shoals or may be considered as a distinguishing component of a particular lithostratigraphic unit over a limited area (McFarlane, 1992; Salad Hersi & Lavoie, 2000). In these cases, their occurrence has been interpreted as an indicator of peritidal to near-shoaling facies in localized regions only (Brown, 1997). Therefore, without significant prominence in stratigraphic sections, they have rarely been recognized with regard to potential event stratigraphic significance.

However, as suggested by Sheehan and Harris (2000) and Riding (2000), the resurgence of stromatolites in post-Cambrian depositional environments is often timed with the loss of diverse marine communities – i.e. after major extinctions. The former authors referred to stromatolites and thrombolites as “opportunistic disaster forms” that were able to regain widespread distribution only during times when animals were effectively removed from marine environments (i.e. post end-Ordovician Extinction, or Early Triassic). Therefore it is not surprising that, during the great Ordovician radiation event of the early Late Ordovician, carbonate environments would be characteristically stromatolite-poor if conditions favored diverse marine communities that would graze on and otherwise disturb bacterial mats. Nonetheless, as recognized in this study their occurrence within relatively narrow depositional windows, tied with other restricted depositional facies across multiple outcrop regions, suggests that their appearance may indeed by indicative of unique events with correlative potential. In essence, these facies might be excellent candidates for time-specific facies in the notion of Walliser (1986).

As with stromatolites, ooid occurrences appear to be highly restricted in the Late Ordovician. To date, ooid formation is still highly debated as per Tucker and Wright (1990), and there are at least 3 different competing hypotheses proposed for the processes responsible for their formation. These include: 1) mechanical, the abiotic “snow-ball” model; 2) chemical abiotic model; radial crystal growth on ooids, and 3) biological model, form due to endolithic and epilithic cyanobacteria that influence mineral growth. Although it is likely that all three mechanisms may be responsible for the formation of some ooid deposits, it is thought that at least for radial and tangential ooids the latter two likely are responsible for most ooid deposits.

As a result of experimental studies (Davies et al., 1978), it has been suggested that radial and tangential ooids are actually produced under slightly different depositional characteristics. In these studies, it was shown that radial forms may grow in low-energy, non-agitated conditions with supersaturated seawater solutions if they contain appreciable concentrations of humic acids. Humic acids are usually formed by the microbial degradation of plant and animal tissues after the death of living cells, and can be retained in soils or other environments where they are not destroyed or rapidly diluted. A second form - tangential ooids could be produced from supersaturated seawater conditions under agitated, high-energy conditions without the presence of organic-derived acids. Nonetheless, growth of these tangential ooids could be stopped under acidic (slight drop in pH) conditions or under increased Mg concentrations, but abiotic crystal growth could be restarted if organic membranes redeveloped on the surface of the grains (Tucker & Wright, 1990). Therefore the implication of ooid formation in these two experimental cases is that most ooid formation in the stratigraphic record may in fact rely on biotic controls whether in restricted-low energy environments or in high-energy conditions in shoaling environments.

Calcification Events in the Chazy Group (CE-1)

In the CBR interval, the appearance of ooids as disseminated components of multiple stratigraphic intervals is again not remarkable in the context of a major stratigraphic event. However the recognition of stratigraphic intervals with facies successions showing evidence for environmental restriction (i.e. evaporitic textures and crystal psuedoforms, obvious ooid-rich intervals, and widespread stromatolite/thrombolites occurrences) across wider region are argued to represent significant events that are useful in correlation. Based on suggestions of Riding (2000) and considering the role of cyanobacteria in both stromatolites/thrombolites as well as in ooid precipitation, these might be considered “cyanobacterial calcification events.” Riding notes that they are prevalent in the Early Ordovician, but herein we report the occurrence of at least one well-developed, relatively short-term cyanobacterial calcification event and a second, less well-developed event that is more restricted in its development, both in the Late Ordovician. Based on Riding's assertions (1993, 2000) these calcification events are linked to temporal changes in seawater chemistry with periods of elevated carbonate saturation relating to one or more of the following: 1) high global temperatures, 2) low sea-level, 3) low skeletal carbonate production, or 4) development of alkalinity pumps from stratified basins (see Riding 1993). Paleogeographic and stratigraphic evidence suggests that all of these conditions may have existed intermittently on the GACB.

The occurrence of stromatolites within the CBR has been recognized for some time. Oxley & Kay (1959) recognized their importance in the diverse communities of the earliest coral reefs in the Chazy Group of New York. More recently Griffing (2000) interpreted the domal to columnar stromatolites in the Upper Crown Point as late-stage components of bioherms. Along with *Solenopora* red-algal crusts, birdseye limestones, mudcracks, and disconformity surfaces, they appear to represent a shallow-water community that developed on top of and in between reefal facies during lowstand restriction. Laminated layered microbiolites and domal stromatolites have also been described from the Michigan Basin (subsurface) where they form in association with ooids and occasionally with anhydrite facies as recognized by Nadon and colleagues (2000). Moreover, further west in Wisconsin and Iowa, Dott (2003) suggested that microbiolites may have been a key factor in stabilization of St. Peter quartz-sandstones in nearshore peritidal environments. In this region, only rare ooids are reported although some are reported from the Dutchtown Formation (along with stromatolites) in the southern Illinois Basin (Alberstadt & Repetski, 1989). South of New York, in central Pennsylvania the Loysburg

Formation “Tiger-Stripe” member is also known to contain stromatolites (Laughrey et al., 2003), but to date no ooids. Further south, in the southern Appalachian region, the equivalent interval does not appear to have numerous or at least reported instances of stromatolites and ooids in this particular interval.

Thus it appears that a small-scale cyanobacterial calcification event may have impacted the northern GACB along the margins of the Grenville Shield and the Transcontinental Arch during the mid-Ashbyan (Crown Point) deposition. At this time cyanobacterial growth became relatively widespread during sea-level lowstand (i.e. during deposition of mid Crown Point). The Michigan Basin region may have become substantially isolated and restricted by regressive events such that ooid formation was also possible on the flanks of the Algonquin Arch although the genesis of these ooids is still not well-documented. Further south, it appears that cyanobacterial structures and ooids were not well-developed and may relate to slightly different environmental conditions in this region. Thus this interval represents CE-1.

Calcification Events in the Black River Group (CE-2)

The largest and most well-developed calcification event (CE-2) occurs in the early Turinian. Stromatolites and microbiolites have been noted from most outcrop regions of the GACB and along the margin of the Sevier Basin in the southern Appalachians. CE-2 occurs between siliciclastic events three and four. Stromatolites in this unit are rarely large and domal, except in limited areas, but typically there is an abundance of oncolites, and laterally linked-types in peritidal facies. These are easily seen in the Gull River of Ontario (Grimwood et al., 1998). Moreover, the middle Pamelia, L-Middle Gull River, McGregor Limestone, the Snyder, the upper Shippensburg Limestone, the Bowen-Hardy Creek interval, and the Ottosee oolite interval are all recognized for the occurrence of fairly widespread ooids. In some cases, these ooids represent upward of 90% of allochems in some units (i.e. Snyder Formation of central Pennsylvania; Laughrey et al., 2003). Studies by Weaver (1992) show that even in relatively siliciclastic-dominated Ottosee Shale of Tennessee, oolites are well-developed. In his study, Weaver showed most ooids in the Ottosee were tangential ooids with diagenetically produced phyllosilicate minerals in some laminae in between calcite laminae. He indicated the physils were produced diagenetically after bacterial coatings trapped illite and chlorite clay minerals very early in the history of the ooids. Across the New York and adjacent Ontario regions, this interval is also well-known for preserved evaporitic crystal textures in fine-grained carbonates of the Pamelia-Gull River interval (Textoris, 1968; McFarland, 1997; Salad Hersi & Lavoie, 2000). These features suggest major evaporation occurred in the Lake Simcoe – Adirondack – Ottawa River region. This may also have occurred in other regions – although there are limited reports suggesting the presence of evaporite signatures elsewhere.

Also distinctive of this particular calcification event is the widespread occurrence of intraclastic conglomerates in the early transgressive phase of this interval. Over wide regions, and especially near prominent topographic high areas (Beauharnois Arch, Adirondack Arch, Marmora Arch, Nittany Arch, Tazewell Arch, and even the Cincinnati Arch) numerous and relatively complex intraclastic to polymictic inter-formational carbonate conglomerates have been noted. Collectively they often occur on the platform in the vicinity of sequence boundary surfaces and in some instances record exhumation and truncation of underlying stratigraphic units down to at least Chazy if not older stratigraphic units (Cooper, 1960; Salad Hersi & Lavoie, 2000). Even in the vicinity of the Tazewell Arch and the Sevier Basin, the upper interval of the Fincastle Conglomerate is well developed and correlative of conglomerates on the platform. In addition to these inter-formational conglomerates, a number of intraformational breccias are also noticed – suggesting erosion and of previously cemented hardgrounds.

Clearly the CE-2 event is related to major tectonic modification of the platform and its marginal areas resulting in exposure and clast generation due to erosion and disturbance and rapid down-slope transport along highly steepened ramps. These conglomerates/breccias may also relate, at least in some fashion, to the restriction of much of the interior of the GACB resulting in an evident modification of sea water chemistry that led to the abundance of stromatolites and oolite deposits. Tobin and colleagues (1996) suggested that early marine calcite cementation was pervasive at this time and may have been mediated by bacterial processes. Thus, by implication their work suggests that some of these intraclasts may relate to bacterial-induced hyper-calcification.

Overall, CE-2 represents a fairly short-lived interval (portions of one depositional sequence) where much of the GACB experienced what Riding (2000) referred to as a cyanobacterial calcification event. The ultimate cause of this event is still somewhat enigmatic. However, as suggested by Riding (2000), it appears that this event may relate to extreme warming, lowering of sea-level, tectonically-induced restriction of the GACB, loss (at least temporarily) of many carbonate secreting organisms, and potential localized anoxia and stratification in the vicinity of the GACB – i.e. the Sevier Basin. Whatever the cause of this event, the effects of it were evidently much reduced during the subsequent depositional sequence.

<u>Syn-sedimentary Deformed Strata aka "Seismites"</u>

An increasing number of convoluted, and possibly seismically disturbed, horizons have been recognized in outcrop in New York, Pennsylvania, Kentucky, and elsewhere outside of the main study region. Evidence for synsedimentary deformation in Ordovician strata has been laid out previously (Pope et al., 1997; Ettensohn et al., 1998; Rast et. al., 1999; Ettensohn et al., 2002b; McLaughlin & Brett, 2002; McLaughlin & Brett, 2004; Brett et al., 2004, Jewell & Ettensohn, 2004) and focused primarily on the Kentucky Jessamine Dome, western Virginia, and some discussions for outcrops in the Mohawk Valley of New York (see Brett et al., 2004). However, an ever increasing number of synsedimentary deformation structures are now being recognized in other areas of the GACB including in some strata predating the Trenton Group where most occurrences have been identified. Most of the pre-Trenton deformed intervals occur in the vicinity of the Sevier Basin in Virginia and Tennessee.

Moreover, the sedimentologic conditions and sequence stratigraphic occurrence have become the focus of several research projects including those mentioned. From these studies it is becoming clear that particular facies are more prone to synsedimentary deformation and that deformation events have local to regional continuity – evidence against simple local synsedimentary deformation-sliding on sloped seafloors. In contrast, the relatively widespread scale of these features within constrained stratigraphic intervals points to episodes of fault movement as a likely trigger for deformation (Ettensohn et al., 1998). Activation of tectonic collision along the Laurentian margin would generate earthquake activity close to developing foreland basins, but also in far-field regions such that synsedimentary deformation of strata could result in response to ground shaking. Correlation of such horizons is not unreasonable especially considering: 1) soft-sediment deformation is more prone to occur in specific lithofacies and portions of depositional sequences that may be correlative on their own, and 2) large magnitude earthquakes have been shown to have far-ranging effects in Holocene depositional settings via liquefaction of sediments up to several hundred kilometers from the epicenter (see Pope et al., 1997).

Occurrence of Seismically-disturbed strata in the GACB?

As shown on **figure 9**, across the GACB, synsedimentary deformation structures, interpreted as seismites, are mostly constrained to the Chatfieldian and are coincident with the Vermontian tectophase of the Taconic Orogeny. Only a few occurrences are reported from the Ashbyan to Turinian interval. The proliferation in the number of soft-sediment deformation structures or “seismites” in the Trenton is likely linked to: 1) the significant impact of Vermontian-phase tectonics across the GACB (as opposed to smaller, localized tectonic activity of the Blountian Phase), 2) somewhat increased slopes in the vicinity of the foreland basin and along developing intracratonic high and low areas, and 3) development of “seismite-prone” facies that are susceptible to shaking and failure and characteristic of Trenton sequences.

In total, for the Trenton and its equivalents, there have been approximately eleven seismite horizons reported from outcrops in the Cincinnati Arch region and at least 6 from the Mohawk Valley of New York State (Brett et al., 2004). In some instances, these deformed intervals have also been recognized in cores as well – especially in the subsurface of Ohio and Kentucky. Outside of these regions, soft-sediment deformation has also been reported and observed from the Salona to lower Coburn formations (Laughrey et al., 2004), and the Hershey (Jacksonburg Limestone) of Pennsylvania. At least 2 different deformed horizons are known from the lower to medial Martinsburg/Dolly Ridge interval in West Virginia to Virginia (Pope et al. 1997). Significant numbers of deformed beds occur in late Martinsburg strata of Virginia as well, but are beyond the scope of this study. Deformed structures are also reported from the lower argillaceous member of the Hermitage Limestone of the Nashville Dome in Tennessee (Holland & Patzkowsky, 1997). In pre-Trenton strata, the number of seismites is significantly less and these appear over a substantially narrower region.

Chazy Soft-Sediment Deformed Intervals

The earliest soft-sediment deformation structures reported from CBRT strata appears to be from the Rockcliffe Formation in the Ottawa Valley with possible equivalents in the base of the Laval Formation in the St. Lawrence Lowlands. Additional synsedimentary deformation has been shown from the uppermost interval of the Beekmantown Group Carillon Formation located immediately below the Chazy (Dix & Al Rodhan, 2006) in the Ottawa-Bonnechere Graben. Collectively, these beds show evidence for dewatering diapirs, flame structures, ball and pillow structures, and localized brecciation within a succession of sandstones and massive-bedded dolomitic limestones (Dix and Molgat, 1998). In adjacent New York State, synsedimentary fault activity has been postulated to explain rapid thickness variations (Bechtel & Mehrtens, 1995), and major channel features in some levels of the Chazy in the type Chazy region (Oxley & Kay, 1959). Intraclastic breccias with fairly large clasts can also be found in some of these channels suggesting there might be localized disruption of bedding. Mehrtens and Selleck (2002) also provided evidence for possible rotation of cemented nodules and development of wavy-disrupted fabrics in "ribbon limestone" facies of the lower Chazy in the Crown Point region. However no "ball & pillow" or soft-sediment deformation structures have yet been reported – at least not on the scale recognized in the Trenton.

Nonetheless, Dix and Molgat (1998), Salad Hersi and Dix (1999), and Dix and Al Rodhan (2006) indicate that the Beauharnois Arch and Ottawa-Bonnechere graben activated during development of the pre-Chazy Knox Unconformity and formed the Ottawa Embayment leading into the early Chazyan. The area transformed into a locally subsiding protected lagoon during the Chazy with the Beauharnois Arch completely restricting the embayment during the early to mid-Turinian forming an evaporitic lagoon that eventually was breached to the southwest across the Frontenac Arch. Thus the occurrence of soft-sediment deformation in the Chazy – Ottawa Embayment region appears to be tied to localized fault activity in the Ottawa-Bonnechere Graben-Beauharnois Arch region at this time and likely reflects distal tectonic activity associated with collision in the Canadian Maritime region further to the northeast.

Black River Soft-Sediment Deformed Intervals

In the southern Appalachians the occurrence of soft-sediment deformation appears in strata in central to southwestern Virginia, eastern Tennessee and northeastern Georgia. Thus soft-sediment deformation appears to be constrained to the region of the Sevier Basin and along the Tazewell Arch. Cooper (1968), and Lowry and Cooper (1970) report the occurrence of a number of "penecontemporaneous downdip slump structures" in limestones from the Tennessee border northward to the Shenandoah Valley. These have been compared to and are considered analogous with soft-sediment deformation structures located in at least two levels within the earliest Turinian-aged Liberty Hall Limestone and Lower Fincastle Conglomerate near Blacksburg Virginia (Pritchard, 1980).

Pritchard (1980) recognized a major soft-sediment deformation interval seventeen meters above the top of the Effna/Benbolt Limestone in the basal Liberty Hall Formation and also recognized soft-sediment deformation within clasts of an intraformational conglomerate nine meters higher. The conglomerate in this case is embedded in dark-gray, peloidal calcisiltstones within a lenticular "channel-fill" structure and is surrounded by pelletal lime mudstone. The beds within the channel were also somewhat cohesive as they have been folded disharmonically and lie in angular discordance with the surrounding strata. These features appear to be nearly identical to many of the "seismite" horizons in the Lexington Formation of Kentucky. Although they are commonly considered to represent down-slope slump deposits, Pritchard (1980) considered both unstable slopes and seismic shock as the likely trigger mechanisms on locally steepened slopes. Nonetheless, as reported by Cooper (1968) and Lowry and Cooper (1970), nearly identical features can be located in somewhat shallow to slightly deeper carbonate facies. These are exposed intermittently in outcrops from the Tennessee border northward along the outcrop belt in Virginia for up to 350 km or more. These authors recognize two distinct zones of disharmonic "soft-rock flowage folds" in the Edinburg Formation which contain exceptionally large blocks of semi-lithified limestones of varying lithologies. Although their correlation with those of Pritchard is not formally established, it is likely that these deformed intervals are nearly age equivalent.

Nonetheless, the occurrence of many of these features in close proximity to the Tazewell Arch, combined with numerous intraformational and extra-formational conglomerates and breccias in the Liberty Hall-Fincastle-Wardell interval, and the appearance of the Ben Hur Limestone sitting directly on Beekmantown strata (with major truncation of intervening strata) suggest that the Tazewell Arch and intervening areas were indeed activated in similar fashion to the Beauharnois Arch. In Virginia, deformed strata appear to flank both sides of the uplifted Tazewell Arch – although substantially greater thicknesses of strata appear on the southeastern flanks of the Tazewell Arch where the Fincastle Conglomerate is best developed. Lowry and Cooper (1970) suggest that the deformed beds in the Edinburg Formation were shed/moved down-slope to the northwest of the Arch, while those recognized by Pritchard (1980) appear to have been shed to the southeast. Combined with greater thicknesses of strata on either side of the Tazewell Arch – it appears that there may have been a distinct but separate platform that formed east of the main GACB – and may have been a northern up-dip extension of the Sevier

Basin.

Farther south, deformed strata have been recognized in Tennessee in the upper Sevier to medial Ottosee Formation. Neuman (1955) recognized soft-sediment “slump” structures at the base of the Bacon Bend Member of the Sevier and another approximately fourteen meters higher. The deformed beds appear in interbedded strata dominated by calcareous shales and thin to medium-bedded sandstones and siltstones. Again, these features show ball and pillow structures and some deformed disc and saucer morphologies along with dewatering structures. Deformed beds appear as lenses within undeformed strata and folds are characteristically disharmonic with no preferential fold axis dominant (Neuman, 1955). Preservation of many contorted laminations suggest these beds may also have had bio-films or cyanobacterial mats that helped to bind the un-cemented sediments.

These particular deformed beds appear to be somewhat younger than the Liberty Hall deformed beds, but appear to be contemporaneous with the onset of the stromatolite and ooid-rich interval (CE-2) on the GACB as previously discussed. The Bacon Bend deformed interval also immediately predates the deposition of the Bays Formation and the major volcanic eruptions of the later Turninian. Nonetheless, the coincidence of extensive fault-induced (?) seismic slumping in the vicinity of the Sevier Basin/Tazewell Arch, and near simultaneous restriction of the GACB suggest a major pulse of tectonic activity in the Blountian Phase of the Taconic Orogeny. This evidently resulted in uplift along the edge of the cratonic margin (thus protecting and restricting interior areas), as well as uplifting areas of the hinterland resulting in rapid infilling of the intervening Sevier Basin as recorded by the rapid change from flysch-style deposition of the Sevier to red bed molasse phase deposits of the Bays Formation.

Subsequent to the Bacon Bend deformed interval – there are very few additional reports of deformed intervals in the literature of the Black River Group and its equivalents. In a few places evidence for synsedimentary deformation has been reported, but is often linked to evaporitic expansion crusts (teepee structures) or storm rip-up intervals. Although in some regions, deformation prone facies were clearly developed as were somewhat steepened ramps, the "seismite" horizons are generally absent until the early Chatfieldian.

Trenton Soft-Sediment Deformed Intervals

The next recorded interval of deformed strata appears in the basal Lexington Group in the uppermost Curdsville to Logana Limestones. These appear in the vicinity of the Jessamine Dome after the Turinian K-bentonite swarm leading up to the Capital K-bentonite and become very pervasive in the vicinity of the Kentucky River Fault System at different levels within the Lexington Limestone (see Brett et al., 2004). Although somewhat less pervasive, synsedimentary deformed intervals are also recorded from the Napanee Limestone of central New York State and at several different horizons in overlying Trenton strata. The number of synsedimentary deformed intervals rapidly increases in medial-to-late Trenton/Lexington strata.

The coincidence of these widespread seismite horizons along with initiation of shale deposition, and rapid regional subsidence (and global eustatic flooding) suggests another major tectonic pulse of the Taconic Orogeny initiated early in the deposition of the Trenton Group during the Vermontian Tectophase. Sometime thereafter, the most rapid phase of tectonic convergence initiated and major over-thrusting of the Laurentian cratonic margin resulted in major widespread destabilization and movement of ancestral faults well into the interior of the GACB. This evidently included portions of the Adirondack Arch, Cincinnati Arch, Sebree Trough, the Algonquin Arch, and perhaps even the Wisconsin Arch and its southern equivalents.

Chemostratigraphic Events

General Introduction and Background

Within the CBRT interval ever increasing attention has been focused on documenting and defining potential geochemical signatures from various outcrop strata around the GACB. The majority of these studies have been focused on stable isotopes including those of carbon, oxygen, neodymium, and strontium. These investigations have not only focused on the recognition and use of isotopic changes for regional correlation (i.e. carbon isotopic excursions), but also have been used to help infer something about the dynamics of paleoenvironmental change in the GACB region during the Late Ordovician. For instance, in their study of Ordovician brachiopod isotopic composition, Shields and colleagues (2003) presented evidence for carbon, oxygen, and strontium isotopic change throughout the Ordovician from ten different worldwide localities.

Using the updated global time-slices proposed by Webby and colleagues (2004), the former authors have recognized significant and substantial short-term isotopic changes in the Middle to Late Ordovician transition and specifically in the 5b to 5c time-slices. These time slices coincide with the Turinian-Chatfieldian interval and the Taconic Orogeny. Shields and colleagues (2003) proposed that the dramatic drop in strontium isotopes may have been related to substantially lower continental erosion rates (owing to global transgression) coupled with increased submarine hydrothermal venting due to increased sea floor spreading rates. For oxygen isotopes, these researchers recognize a substantial rise in $\delta^{18}O$ through the same interval and suggest that it is related to cooling of tropical seawaters prior to the end Ordovician glaciation. As for carbon, Shields and colleagues (2003) recognize a relatively long-term gradual climb in $\delta^{13}C$ that is punctuated by the end Ordovician positive carbon isotopic excursion, as well as several other important excursions in the lower Upper Ordovician. This transition has been attributed to increasing carbon burial rates due to increased tectonic activity and uplift.

Collectively Shields et al. data suggest that the relatively synchronous change in these three isotopic systems (including the largest recorded swings in both $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ of the Phanerozoic), combined with one of the largest marine transgressions in the Phanerozoic, argue for a major reorganization of ocean chemistry and surface environments during the 5a, 5b, and 5c time slices of Webby and colleagues (2004). Unfortunately, as of yet, the resolution of these isotopic curves is lacking and awaits further refinement. Nonetheless major progress has been made toward refinement of the carbon isotopic curve and recognition of several distinct excursions in various parts of the GACB. Herein the status of major isotopic excursions, thus far recognized, is reviewed. Moreover, this study also integrates strontium isotope data from the literature in an effort to produce a more-refined strontium isotopic curve for the CBRT interval which is calibrated against the numerous event horizons and stratigraphic intervals used in this study.

Carbon Isotopes:

<u>General Introduction & Carbon Cycling:</u>

Shields and colleagues (2003) identified the Late Ordovician as an interval of relatively substantial and synchronous change with respect to multiple isotopic systems including carbon, oxygen, and strontium. These authors, using the global time-slices of Webby et al. (2004), specifically identified the mid-Mohawkian (time-slices 5a and 5b) as representing a significant interval of change in global isotopic cycling and attributed these to changes in the equilibrium of the global ocean as a result of tectonics, sea-level change, and climatic change. However, the isotopic shifts in the 5a-5b interval predate the end Ordovician excursions associated with the

Hirnantian Glaciation especially in regard to carbon isotopes (Kump et al., 1999; Finney et al., 1999) and are a subject of significant research and debate.

Overall, as recognized by Shields and colleagues (2003) $\delta^{13}C$ values of carbonate materials show a transition from slightly negative (lighter) $\delta^{13}C$ values in the Early Ordovician to slightly positive (heavier) $\delta^{13}C$ values in the later Ordovician. This long term trend is thought to be associated with the removal of ^{12}C (the lighter stable carbon isotope) from the Ordovician ocean and atmosphere reservoirs most likely by biological fractionation and burial of organic matter. After removal of ^{12}C, remaining seawater is preferentially enriched in the heavier ^{13}C isotope. Thus when carbonates are produced from enriched seawater $\delta^{13}C$ values for carbonates are in turn enriched (heavy) and buried $\delta^{13}C$ of the organic fraction is likewise depleted or light, but also gradually become heavier.

The flux of stable carbon isotope components between reservoirs (i.e. the atmosphere and the oceans) varies as a function of the rate of input of carbon from weathering of rocks and volcanic degassing processes as well as rates of sediment burial either as organic matter or as precipitation and burial of carbonate minerals (Kump and Arthur, 1999; Fanton and Holmden, 2007). Thus in most instances in the geologic record, changes in $\delta^{13}C$ values of carbonates, if positive, generally reflect extensive removal of organic carbon from circulation in the global ocean (usually with $\delta^{13}C$ ranging from 2 to ~7‰ for carbonate fractions). This could occur either during ocean stagnation and anoxia (lower ocean water masses retain significant light carbon with negative $\delta^{13}C$ values ~ -29‰), or during periods where sedimentation rates are high and burial occurs rapidly regardless of oceanic stagnation or circulation.

Although much less likely to affect global oceanic $\delta^{13}C$, slight increases in $\delta^{13}C$ values to near 0‰ in local to regional areas (i.e. partially restricted basins & seas) could also be related to the increased flux, via fresh water runoff, of bicarbonate from weathering of silicate rocks and/or dissolution of exposed carbonate rocks (Kump & Arthur, 1999; Fanton & Holmden, 2007). In contrast, negative shifts in $\delta^{13}C$ values are commonly attributed to the re-introduction of light carbon (^{12}C) to circulating reservoirs. The return of light carbon can occur either by re-invigoration of ocean circulation (i.e. ending stagnation), volcanic degassing ($\delta^{13}C$ values ~-5-10‰), weathering oxidation of previously deposited organic-rich rocks ($\delta^{13}C$ values ~-22‰), or the release of biogenic methane from the seafloor ($\delta^{13}C$ values averaging -50‰) (Kump and Arthur, 1999).

Arthur and Kump (2002) suggested that, based on theoretical modeling, it is very difficult to produce large positive carbon isotope excursions and thus their appearance in the rock record indicates significant perturbations to the cycling of carbon. These authors reiterated that negative carbon isotope excursions can be produced by episodes of methane release, volcanic eruptions, or increased rates of mid-ocean ridge basalt degassing with lowered rates of biological fractionation. Thus understanding the dynamics of the global carbon cycle is very important to understanding the nature of the Ordovician carbon isotopic signals, moreover it is also important to understand regional geography to understand the potential for local to regional variations in carbon cycling due to partial restriction and the potential for nutrient loading. Although the Ordovician certainly witnessed a long-term change in carbon isotopes owing to at least several of the previously outlined scenarios, it is now apparent that the long-term pattern of isotopic change is also punctuated by an increasing number of shorter-term events. These may reflect an important high-frequency oscillation and likely relate to sea-level or climate oscillations superimposed on any large-scale tectonic influences.

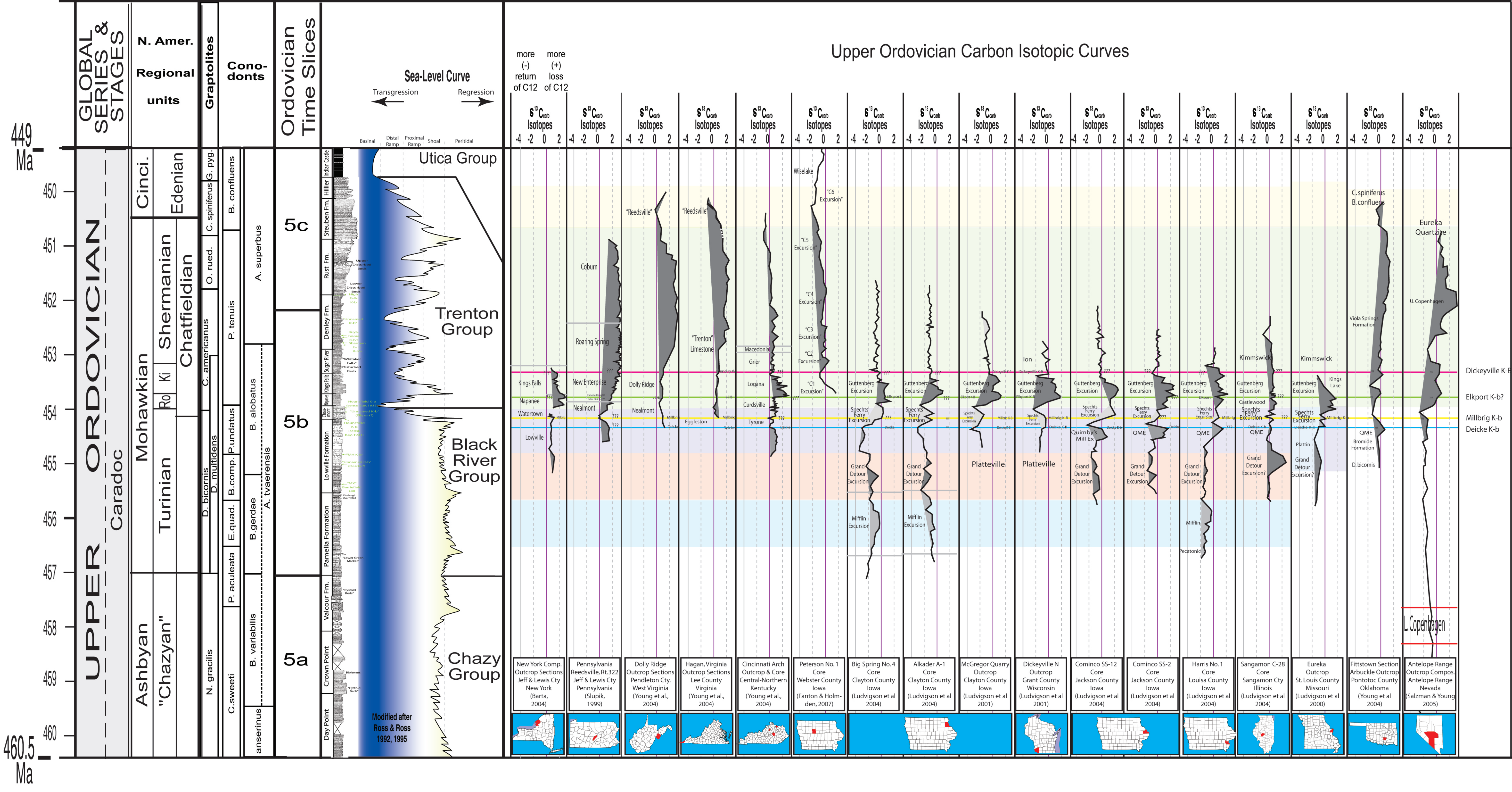
Ordovician Carbon Isotope Excursions: GICE

Upper Ordovician strata of the GACB interval are now recognized to have at least two large-scale carbon isotopic shifts and a number of smaller-scale changes. The best-known large-scale positive excursion includes the end-Ordovician excursion ($\delta^{13}C$ ~7‰) recognized during the Hirnantian glacial lowstand (Kump & Arthur, 1999; Kump et al. 1999; Saltzman & Young, 2005). The oldest larger-scale event, although somewhat smaller in scale than the Hirnantian excursion (max $\delta^{13}C$ ~4‰), has been recognized from the lower Chatfieldian stage of the Upper Mississippi Valley. The excursion begins immediately above the Millbrig K-bentonite and extends upward to just below the position of the Dickeyville K-bentonite (Ludvigson et al., 2000) (**figure 17**). This event has been referred to as the "mid-Caradoc Carbon Isotopic Excursion" or more commonly as the "Guttenberg Isotopic Carbon Excursion" (or GICE). This particular event appears to represent a global event as it has been equated with similar excursions in the Baltic region and China (Ainsaar et al., 1999; Saltzman et al., 2003).

The GICE is recognized in the Upper Mississippi Valley in outcrops and cores as a slightly positive $\delta^{13}C$ excursion of 1-2 ‰ above pre excursion values which are typically less than 0‰. In this region, the maximum amplitude of the GICE is somewhat variable, but $\delta^{13}C$ values typically range up to 2‰. As suggested by Ludvigson and colleagues (2004), GICE values show slightly heavier values overall toward the south away from the Hollandale Embayment and the Transcontinental Arch. Fanton and Holmden (2007) suggested the increase in $\delta^{13}C$ values related to proximity to the intracratonic Sebree Trough which was likely a source for upwelling of nutrients and increased organic carbon burial.

Other studies have worked to demonstrate the distribution and occurrence of the mid-Caradoc GICE $\delta^{13}C$ excursion in other areas of the GACB region (Fanton & Holmden, 2001; 2007; Ludvigson et al., 1996, 2000, 2002; 2004; Ludvigson and Witzke, 2005; Panchuk et al.,

Figure 17: Carbon isotopic curves for the Turinian-Chatfieldian strata across the GACB region including New York, Pennsylvania, West Virginia, Virginia, Iowa, Wisconsin, Illinois, Missouri, and Oklahoma. The stratigraphic position of profiles is established relative to the position of the four K-bentonites: Deicke, Millbrig, Elkport, and Dickeyville as currently established. Data is compiled from reports as shown and include data presented by: Barta (2004), Fanton & Holmden (2007), Ludvigson et al., (2000, 2001, 2004), Salzman & Young, (2005), Slupik (1999), Young et al., (2004)



Upper Ordovician Carbon Isotopic Curves
GLOBAL SERIES & STAGES
N. Amer. Regional units
Graptolites
Cono-donts
Ordovician Time Slices
Sea-Level Curve
Transgression
Regression
Basinal
Distal Ramp
Proximal Ramp
Shoal
Peritidal
more (-) return of C12
more (+) loss of C12
δ13Ccarb Isotopes
-4 -2 0 2
449 Ma
450
451
452
453
454
455
456
457
458
459
460
460.5 Ma
UPPER ORDOVICIAN
Caradoc
Mohawkian
Ashbyan
"Chazyan"
Turinian
Shermanian
Cinci.
Edenian
Chatfieldian
Ki
Ro
G. pyg.
C. spiniferus
O. rued.
C. americanus
D. multidens
D. bicornis
N. gracilis
B. confluens
P. tenuis
P. undatus
B.comp.
E.quad.
P. aculeata
C.sweeti
A. superbus
B. alobatus
B.gerdae
A. tvaerensis
B. variabilis
anserinus
5c
5b
5a
Utica Group
Trenton Group
Black River Group
Chazy Group
Modified after Ross & Ross 1992, 1995
Indian Castle
Hillier
Steuben Fm.
Rust Fm.
Denley Fm.
Sugar River
Kings Falls
Napanee
Chaumont
Lowville Formation
Pamelia Formation
Valcour Fm.
Crown Point
Day Point
Dickeyville K-B
Elkport K-b?
Millbrig K-b
Deicke K-b
Kings Falls
Napanee
Watertown
Lowville
Coburn
Roaring Spring
New Enterprise
Nealmont
"Reedsville"
Dolly Ridge
Nealmont
"Trenton" Limestone
Eggleston
Macedonia
Grier
Logana
Curdsville
Tyrone
Wiselake
"C6 Excursion"
"C5 Excursion"
"C4 Excursion"
"C3 Excursion"
"C2 Excursion"
"C1 Excursion"
Guttenberg Excursion
Spechts Ferry Excursion
Grand Detour Excursion
Mifflin Excursion
Platteville
Ion
Quimby's Mill Ex
QME
Mifflin
Pecatonic
Kimmswick
Castlewood
Grand Detour Excursion?
Plattin
Kings Lake
C.spiniferus
B.confluens
Viola Springs Formation
Bromide Formation
D.bicornis
Eureka Quartzite
U.Copenhagen
L.Copenhagen
New York Comp. Outcrop Sections Jeff & Lewis Cty New York (Barta, 2004)
Pennsylvania Reedsville, Rt.322 Jeff & Lewis Cty Pennsylvania (Slupik, 1999)
Dolly Ridge Outcrop Sections Pendleton Cty. West Virginia (Young et al., 2004)
Hagan, Virginia Outcrop Sections Lee County Virginia (Young et al., 2004)
Cincinnati Arch Outcrop & Core Central-Northern Kentucky (Young et al., 2004)
Peterson No. 1 Core Webster County Iowa (Fanton & Holmden, 2007)
Big Spring No.4 Core Clayton County Iowa (Ludvigson et al 2004)
Alkader A-1 Core Clayton County Iowa (Ludvigson et al 2004)
McGregor Quarry Outcrop Clayton County Iowa (Ludvigson et al 2001)
Dickeyville N Outcrop Grant County Wisconsin (Ludvigson et al 2001)
Cominco SS-12 Core Jackson County Iowa (Ludvigson et al 2004)
Cominco SS-2 Core Jackson County Iowa (Ludvigson et al 2004)
Harris No. 1 Core Louisa County Iowa (Ludvigson et al 2004)
Sangamon C-28 Core Sangamon Cty Illinois (Ludvigson et al 2004)
Eureka Outcrop St. Louis County Missouri (Ludvigson et al 2000)
Fittstown Section Arbuckle Outcrop Pontotoc County Oklahoma (Young et al 2004)
Antelope Range Outcrop Compos. Antelope Range Nevada (Salzman & Young 2005)

2006; Saltzman et al., 2001; Saltzman & Young, 2005; Young et al. 2005). The original "mid-Caradoc isotope excursion" or GICE, was defined in the upper Mississippi Valley region for outcrops and cores in Iowa, and Wisconsin. Nonetheless, this excursion is now recognized as the uppermost of five Turinian to early Chatfieldian excursions correlated in the Wisconsin Arch to Illinois Basin region (Wisconsin, Iowa, Illinois, and Missouri). The occurrence of these additional excursions indicates an increasingly complex scenario for carbon cycling in the GACB (Ludvigson et al., 2004; Ludvigson and Witzke, 2005). More recently, immediately above the level of the GICE, Fanton and Holmden (2007) identified another five $\delta^{13}C$ excursions ranging upward to the base of the Wise Lake Formation (lower Edenian stage). Thus at least ten positive $\delta^{13}C$ excursions are represented in Turinian to Chatfieldian strata.

More regionally in the GACB, the GICE excursion has been extended to Kentucky, Virginia, West Virginia, and New York using the position of key K-bentonites and the relative position of the *P. tenuis* conodont biozone boundary (Young et al., 2005; Barta, 2004). The GICE has also been tentatively correlated into the Arbuckle region of Oklahoma and into Nevada using biostratigraphy (Young et al., 2005; Saltzman & Young, 2005). Likewise in central Pennsylvania, an excursion was also recognized previously in the Salona and Coburn Formations relative to the position of two K-bentonites thought to be the Millbrig and Deicke K-bentonites (Patzkowsky et al., 1997). The accuracy of these K-bentonite correlations is problematic and has been contended on grounds documented herein.

Nonetheless, in the central Pennsylvania to West Virginia region, $\delta^{13}C$ values approach the highest values of the Mohawkian ($\delta^{13}C$ values of up to 4‰). In Kentucky, Virginia, and West Virginia $\delta^{13}C$ values typically range from 0 to 2‰ in pre-GICE strata and rise to over 3‰ in post GICE strata. Thus, along the margins of the GACB, in the vicinity of the Sebree Trough, and in the Taconic Foreland (central Pennsylvania, Virginia, and West Virginia), the $\delta^{13}C$ values of carbonates are heavier than values in near-shore regions of the GACB. Regional $\delta^{13}C$ values have been mapped out into distinct aquafacies zones (Young et al., 2005; Panchuk et al., 2006) with the "Taconic Aquafacies" generally showing a much more enriched and prolonged duration for the excursion compared to GACB interior regions.

Biostratigraphic data from Virginia and West Virginia suggest the excursion ranges through the *P. tenuis* conodont chronozone into the *B. confluens* zone near the top of the Chatfieldian before returning to lighter values. The GICE in the Upper Mississippi Valley returns to lighter values significantly earlier than the *B. confluens* biozone boundary. Thus although the GICE apparently initiated at roughly the same time on the GACB as the mid-Caradoc excursion within the foreland basin, the longer duration of the excursion in the foreland indicates that these areas were hydrodynamically decoupled from shallower portions of the GACB. Moreover, with additional carbon isotopic excursions recognized from Chatfieldian strata above the level of the GICE (Fanton & Holmden, 2007), it is likely that the extended excursion of the foreland may actually represent a composite excursion. In this scenario each discrete excursion recognized on the platform is stacked into a single longer-term isotopic excursion in foreland basin areas. The compositing of successive excursions makes it difficult to differentiate the distinct excursions recognized elsewhere.

Ordovician Carbon Isotope Excursions: PRE-GICE Positive Excursions

In addition to the GICE, Ludvigson and colleagues (2004) and Ludvigson and Witzke (2005) have recognized instability in the carbon isotopic record of carbonates below the position of the GICE within Turinian strata of Iowa and Wisconsin. Collectively, Ludvigson and colleagues (2004) have recognized five positive $\delta^{13}C$ excursions in the Turinian to lower

Chatfieldian interval (**see figure 17**). These excursions have significantly lower amplitudes compared to the GICE and the end-Ordovician excursion; however, they are recognized in several core and outcrop regions and therefore appear to be robust signals. These have been named the Mifflin, Grand Detour, Quimby's Mill (all below the Deicke K-bentonite), the Spechts Ferry (between the Deicke and Millbrig K-bentonites), and the Guttenberg (GICE) excursion (located above the Millbrig and centered about the level of the Elkport K-bentonite). These shifts are superimposed on the long-term Ordovician trend toward heavier $\delta^{13}C$ values, and have been attributed to sea-level changes in the GACB epeiric sea (Fanton & Holmden, 2005; Panchuk et al., 2006).

Outside of the Iowa region, pre-Chatfield positive $\delta^{13}C$ excursions have not been effectively documented for correlation with those recognized by Ludvigson and colleagues (2004). However, data presented by Young and colleagues (2005) from central Kentucky shows some evidence for the Spechts Ferry Excursion in the upper Tyrone to lower Curdsville Formation between the Deicke and the Elkport K-bentonite or its probable equivalent (Capitol). In central Pennsylvania, below the level of the pronounced, long-duration excursion which occurs in the Salona Formation, another positive $\delta^{13}C$ excursion is also noted in the Nealmont Formation (Slupik, 1998). If K-bentonite correlations have indeed been misinterpreted in this region, this Nealmont excursion could be equivalent to the Spechts Ferry Excursion elsewhere. Below the level of the Deicke K-bentonite Young and colleagues (2005) indicate the presence of at least one positive excursion in the High Bridge Group of central Kentucky. This excursion appears to be coincident with the Quimby's Mill excursion or the subjacent Grand Detour excursion. Likewise data from Barta (2004) indicates the presence of a diminished excursion in the Lowville Formation well below the Millbrig that may correlate with either of these excursions. In the upper Bromide of the Arbuckle region, Young and colleagues (2005) recognize another small positive excursion that could correlate with either the Grand Detour or the Quimby's Mill excursion based on biostratigraphy. Thus, clearly, significant work must be done before confidently correlating these lower excursions. However these preliminary data suggest the possibility for more regional correlation of these excursions.

<u>Ordovician Carbon Isotope Excursions: POST-GICE Positive Excursions</u>

In addition to the Turinian excursions, Fanton and Holmden (2007) recognized five positive $\delta^{13}C$ isotopic excursions in strata overlying the GICE. Again, these are diminished in their amplitude compared to the GICE. However these excursions are recognized in carbonate fraction, organic carbon fraction and in the total organic carbon values (by weight percent). Four of these excursions have been recognized in the Dunleith Formation, and one in the Wise Lake Formation. Starting with the GICE, these are referred to as the C1 (GICE), C2, C3, C4, C5 and C6 excursions. As with the pre-GICE excursions, these have not been correlated outside of the Iowa region. Nonetheless, their appearance can be compared to isotopic curves produced for central Pennsylvania (Patzkowsky, 1997; Slupik, 1998), Kentucky, Virginia, and West Virginia (Young et al., 2005). In these latter regions, previous interpretations suggested that these excursions were entirely time-equivalent to the GICE but significantly more heavy ($\delta^{13}C$ values are typically in excess of 2‰). Despite the absolute differences, there are variations in $\delta^{13}C$ values throughout the excursion (both in carbonate and organic fractions) comparable in scale to those recognized by Fanton and Holmden (2007). Combined with available biostratigraphic data, and K-bentonite correlations from the Virginia and West Virginia localities, this evidence suggests that the excursion in the Taconic Foreland is of a much longer duration. Thus although specific speculation about correlation of the C2 through C5 excursions out of Iowa is premature – there is potential for subsequent analysis to investigate possible connections.

One correlation that is potentially viable is the C5-C6 boundary excursion, as it is roughly coeval with the *B. confluens* conodont biozone. As shown in West Virginia, and Virginia, $\delta^{13}C$ values begin to lighten and drop below 0‰ for the first time since the beginning of the Chatfieldian. In these areas, the lower values are recorded at the base of the "Reedsville" Formation (sandstones and shales) before values once again begin to climb in the medial to upper Reedsville. In central Pennsylvania, although values don't quite reach the 0‰ value before the data ends– they begin to drop to the lowest values just below the level of the Antes Shale in the uppermost Coburn Formation. This decline appears to be coincident with the decline recorded at the end of the C5 excursion and just prior to the initiation of the C6 excursion in earliest Edenian. A similar pattern is observed in the uppermost Viola Springs of the Arbuckle region where $\delta^{13}C$ values of carbonates drop below 0‰ near the base of the *C. spiniferus*, *B. confluens* biozone boundary. Saltzman and Young (2005) note the same pattern at the base of the Eureka Quartzite from Nevada – which they interpret as a major sea-level regression.

Significance of Positive Carbon Isotopic Excursions

In a previous investigation of neodymium isotopes, it was suggested that changes in neodymium isotopes were reflective of high-order sea-level fluctuation (Fanton et al., 2002). As the small-scale positive $\delta^{13}C$ isotopic excursions recognized by Fanton and Holmden (2007) appear to co-vary with the neodymium excursions as previously recognized, these authors suggest that the behavior of the carbon isotopic system in epeiric seas is a function of sea-level forcing and proximity to paleo-shorelines. In this scenario, positive $\delta^{13}C$ excursions resulted from higher rates of organic carbon burial during and after marine flooding events. Subsequently during sea-level lowstand, strata record lighter $\delta^{13}C$ values as the influx of light carbon increases from weathering of exposed deposits and increased thermo-haline circulation increases upwelling and mixing of isotopically lighter deep water (Fanton & Holmden, 2001; Ludvigson & Witzke, 2005; Fanton & Holmden, 2007).

The larger-scale GICE, due to its stratigraphic position relative to other key marker beds, has been fairly well documented. Its initiation is coeval with the first sea-level rise event of the Trenton. However, the cause of the pronounced excursion (the GICE) is still debated. The initial phase of the excursion appears to be correlative and isochronous between Iowa, Oklahoma, Kentucky, Virginia, West Virginia, Pennsylvania, Ontario, and New York (Patzkowsky et al., 1997; Bergström et al., 2001; Barta 2004; Young et al., 2005). In the latter two regions – the GICE reaches its maximum $\delta^{13}C$ values (between 2.0 and 3.0 ‰) within the type Rockland strata (Selby-Napanee Formations) before dropping back to lighter values in the shallowing transition into the overlying Kings Falls Formation. This pattern is nearly identical to that seen in strata of central Kentucky, and the Upper Mississippi Valley. In central Pennsylvania, Virginia, and West Virginia, the heaviest $\delta^{13}C$ values approach 4 ‰ and are sustained throughout the rest of the Chatfieldian before dropping back to lighter values in post Chatfieldian strata. The intensification of $\delta^{13}C$-enrichment of the foredeep areas is likely due to isolation of the platform from the foreland basin in terms of paleoceanographic circulation and greater biological fractionation. The lack of access to less-fractionated (lighter) water masses, would explain the more drawn out “GICE” excursion as recognized in Oklahoma, Nevada, Virginia, and West Virginia (Young et al., 2005), or the “mid-Caradoc isotopic carbon excursion” in central Pennsylvania (Patzkowsky et al., 1997; Slupik, 1998).

Moreover, still not entirely clear is why the GICE overall has a much larger magnitude relative to subjacent and superjacent excursions if they are all entirely tied to sea-level change. Thus, additional implications for the GICE excursion need to be considered. It was proposed previously that the "mid-Caradoc" positive $\delta^{13}C$ isotopic excursion was produced as a result of a relatively short-lived, global cooling event predating the end-Ordovician glaciation (proposed by Patzkowsky et al., 1997; and supported by Saltzman & Young, 2005). In this scenario, it was suggested that the onset of glaciation in Gondwana resulted in intense circulation of the southern Ordovician ocean. This meant that Laurentia experienced vigorous nutrient-rich upwelling through the newly activated Sebree Trough, which resulted in significant biologic fractionation of carbon and simultaneous mass-burial of organic carbon at the onset of major Taconic orogenesis (Patzkowsky, 1997; Saltzman & Young, 2005). Saltzman and Young (2005) further proposed that this biologic fractionation episode may have been enough to drop atmospheric pCO_2 levels enough to further enhance icehouse cooling by the Edenian coincident with the deposition of the Eureka Quartzite.

However, as indicated by Patzkowsky and colleagues (1997) the scenario of positive $\delta^{13}C$ excursions occurring during sea-level transgression and highstand (i.e. GICE) is more typical of greenhouse climates and runs counter to the pattern witnessed for the Hirnantian when maximum positive excursions coincide with sea-level lowstand and invigorated thermo-haline circulation. As suggested from sedimentary evidence presented by previous workers (Holland & Patzkowsky, 1996; 1998, Kolata et al., 2001), it is possible that the Sebree Trough played a role in nutrient loading into the mid-continent region. However, the scale of nutrient-loading through this conduit is problematic and does not account for regional variation in the amplitude of the excursion and its longer duration outside of the intracratonic area.

First, the correlation of $\delta^{13}C$ positive excursions with periods of global sea-level rise suggests warming of climates, not cooling. Warming would tend to decrease vigorous thermo-haline driven upwelling and limit the flow of nutrient-rich deep waters to the ocean. It would also favor partitioning of oceanic water masses resulting in heavier $\delta^{13}C$ values in shallower water masses and lighter $\delta^{13}C$ values in deeper water. Without thermo-haline mixing, $\delta^{13}C$ values would become substantially more distinct. Nonetheless intensification of surface weather patterns in sub-tropical latitudes during warming may be enough to induce cooler, nutrient-rich upwelling in shallower regions of the GACB. This scenario is still problematic in the Taconic Foreland region as it does not explain the increased and prolonged fractionation and heavier $\delta^{13}C$ values recorded in this region.

Paleosalinity studies and paleocurrent analysis from the foreland basin suggest the Taconic foredeep may have been fed at depth by warm, high-salinity (up to 38 ‰) currents (Railsback, 1989; et al., 1989; Railsback et al., 1990a,b). These southwesterly flowing currents were overlain by cooler, lower salinity surface waters that may have had brackish salinities as low as 26-27‰ (Railsback & Anderson, 1989). These observations suggest that deep waters of the Taconic foreland basin were produced and transported from lower latitudes that may have had higher productivity, higher biological fractionation, and therefore produced heavier $\delta^{13}C$ values. The lower salinity surface waters indicate possible freshwater influx as suggested by the Fanton & Holmden (2007) model. In this scenario, freshwater runoff from the Transcontinental Arch, Canadian Shield, and even potentially the Taconic Highlands may have helped to produce a pronounced pycnocline that effectively isolated water masses. This isolation may have resulted in the numerous aquafacies belts identified by Panchuk and colleagues (2005) and Young and colleagues (2005). Moreover given that the record of neodymium isotopes across the GACB at this time yields Taconic-derived signatures, the elevated $\delta^{13}C$ values especially within the Taconic Foreland suggest some nutrients may have originated from Taconic orogenesis. More distant portions of the GACB thus would have had relatively lower nutrient loading and therefore lower biologic fractionation – even when compared to areas fed by the Sebree Trough.

However, the presence of the GICE is additionally enigmatic as it coincides with the initiation of rapid tectonic uplift of distal slope and rise sedimentary rocks in hinterland areas of the Taconic Orogen. This uplift should have initiated weathering of older, metamorphic rocks and organic-rich deposits (with light $\delta^{13}C$ values). This was also a time associated with increased volcanic activity as recorded by voluminous volcanic eruptions during this time. All of these scenarios would tend to favor lighter not heavier $\delta^{13}C$ values (Kump & Arthur, 1999) as are recorded by the GICE. Therefore, in addition to the problematic sea-level rise (during supposed cooling?), any positive isotopic shift would have needed to overwhelm the flux of light carbon from volcanic/weathering sources.

Thus if the primary method for generating a positive $\delta^{13}C$ excursion lies within biological fractionation and burial, the scale of nutrient-loading, both from Taconic sedimentation and Sebree Trough upwelling, must have been immense and of relatively short-duration. Thus it is likely that paleoclimatic conditions, that favored high nutrient delivery rates and increased freshwater influx to the Laurentian region, changed drastically and coincidentally with the onset of the Vermontian phase of the Taconic Orogeny. Perhaps tectonic uplift initiated a change to monsoonal climates – which not only increased the delivery of nutrients to promote increased biological productivity, but also introduced substantial sediment loads that could ensure partitioning of stable carbon isotopes via burial of organic carbon in off-shore areas. This scenario would help explain heavier $\delta^{13}C$ values of the foredeep in regions proximal to the orogen during the Chatfieldian. The return to lighter $\delta^{13}C$ values on the GACB, relative to the foreland basin, during the Kirkfieldian was likely achieved as a result of physical partitioning of the GACB from the foredeep through activation of ancestral fault systems which produced uplift along the Adirondack Arch and its southern equivalents. Thus for the remainder of the Chatfieldian, positive isotopic excursions within the GACB (Fanton & Holmden; 2007 type), were governed primarily by sea-level rise and fall and nutrient upwelling through the Sebree Trough. As mentioned, by the end of the Chatfieldian, the positive $\delta^{13}C$ values across the GACB and in the Foreland Basin drop below 0‰ at roughly the same time leading into the lower Edenian. This change is coincident with the phase of cooling proposed by Saltzman and Young (2005) and sea-level lowering witnessed globally near the end of the Trenton. Although continued dark-shale deposition persisted for some time thereafter in portions of the foredeep as it migrated cratonward, substantial shallowing and coarsening of facies into flysch phase deposition indicates that major tectonic uplift had subsided and climates may have indeed begun to cool as suggested by previous authors.

Ordovician Carbon Isotope Excursions: Negative Excursions?

Although previous studies have focused primarily on the positive isotopic carbon excursions recorded from the Turinian-Chatfieldian interval, no emphasis has been placed on the negative portions of the published isotopic carbon curves. As shown in **figure 17** each positive $\delta^{13}C$ excursion is usually separated by a pronounced negative and occasionally sharply negative excursion. The nature of these negative excursions is complicated by the nature of the stratigraphic intervals within which they commonly occur. As suggested by Fanton and Holmden (2007), the majority of the lighter carbon isotopic shifts appear to coincide with sea-level lowstand events. Therefore, in some areas, especially where strata become emergent, the full extent of the negative isotopic excursions may be unrecorded or alternatively eroded during the lowstand event itself. In the Fanton and Holmden study (2007) the slight negative excursions in the Chatfieldian are typically less pronounced when compared to those of the pre GICE interval. For instance, in the Clayton County, Iowa cores $\delta^{13}C$ values of carbonates approach values of -3‰ in Turinian strata. Elsewhere in this portion of the Turinian, negative values range normally between -2 to -1‰. In the Chatfieldian the negative excursions that define the boundaries between the C2-C5 positive excursions are typically between 0 to -2‰. Not until the C5-C6 boundary excursion do values dip again to levels of the uppermost Turinian.

Given the carbon-isotopic fractionation parameters outlined by Kump and Arthur (1999), the -1 to -3 ‰ values dip below values possible from normal river flux of carbon from weathering of silicate and carbonate rocks (rivers average about 0‰) during lowstand as proposed by the Fanton and Holmden model. Therefore it is likely that at least some of these negative excursions need an additional supply of light carbon. During lowstand this could come from weathering/oxidation/burning of organic-rich sedimentary deposits (with average values of -22‰) or increased oceanic mixing via enhanced thermo-haline circulation. When lowered sea-level is difficult to support, negative excursions could result from increased rates of volcanic degassing and/or increased rates of mid-ocean ridge basalt degassing as outlined previously (Kump & Arthur, 1999).

The most pronounced negative $\delta^{13}C$ values appear in the vicinity of the Deicke K-bentonite and bracket the Quimby's Mill positive excursion. $\delta^{13}C$ values approach -4‰ immediately following the Grand Detour Excursion in northeastern Iowa. In southeastern Iowa, and Missouri, the lightest values are more typically between -1.5 and -2‰. In central Iowa, and elsewhere, there is a pronounced negative $\delta^{13}C$ excursion above the Quimby's Mill Excursion (-3 to –2 ‰) that separates it from the overlying Spechts Ferry positive excursion. Thus defined these negative excursions (the most negative of all of the Turinian-Chatfieldian values) are located in the stratigraphic interval above the Ocoonita K-bentonite and range upward to just above the Deicke K-bentonite before values become distinctively positive again during deposition of the Spechts Ferry. Strata in this interval typically show an overall deepening upward pattern although they remain in peritidal to shallow sub-tidal environments. In central Kentucky, values are barely negative (-.5 to 0 ‰) and in New York values are characteristically lighter, up to ~(+).5‰. The variability of these negative excursions again appears to follow an onshore-offshore pattern with slightly more positive values recorded in more offshore regions and slightly more negative values in near-shore regions.

Thus, it is readily apparent from the published curves that the stratigraphic intervals within which the Deicke and Millbrig occur generally record the most negative $\delta^{13}C$ values of the entire Turinian-Chatfieldian succession. Although these values are obtained within some of the deepest water facies of the Turinian (although still shallow), the appearance of negative excursions suggests that isotopic carbon values resulted from processes that introduced a significant flux of isotopically light carbon and or resulted in the decline of primary productivity. This latter assessment is supported by analysis of preserved organic materials from the upper Mississippi Valley. Pancost and colleagues (1998) report the demise of photoautotrophic bacteria (including *Gloeocapsomorpha prisca*) at the same time as a rise in green sulfur bacteria. Collectively this suggests development of photic zone anoxia and explains the development of the shale-rich Spechts Ferry interval. Thus the prevalence of volcanic activity and likely seafloor spreading may have had widespread effects on carbon cycling across the GACB – which combined with relative partitioning of the GACB by local sedimentary dynamics and tectonic-induced uplifts to create variable magnitude excursions.

Strontium Isotopes:

General Introduction to Ordovician Strontium Isotopes

Decade old investigations of pristine, well-preserved calcitic and apatitic fossils and primary depositional cements from Cambro-Ordovician strata have helped to produce a reliable range of $^{87}Sr/^{86}Sr$ values for seawater during this time (Denison et al., 1998; Qing et al., 1998). These data have been supplemented by additional recent studies by Shields and colleagues (2003). Collectively these reports, although poorly resolved in terms of regional chronology, have recognized a substantial long-term decline in the $^{87}Sr/^{86}Sr$ isotopic ratio during the Ordovician from maximum values (.7092) in the latest Cambrian to minimum values (.7078) in the Upper Ordovician with minimum values reached within the Caradoc. As pointed out by Young and colleagues (2004) as well as Shields and colleagues (2003), the 0.0014 drop in $^{87}Sr/^{86}Sr$ during this interval is dramatic and in need of greater resolution, but points to very significant changes in the cycling of strontium isotopes in the Ordovician oceans. Nonetheless, the present study provides an excellent opportunity to further refine a $^{87}Sr/^{86}Sr$ curve for the CBRT interval by using other stratigraphic tie lines to assist in placing $^{87}Sr/^{86}Sr$ isotope values into a stratigraphic framework.

Strontium Isotope Stratigraphy and Primary Controls on $^{87}Sr/^{86}Sr$

As pointed out by McArthur (1998), strontium isotope stratigraphy offers an excellent method for correlation and even relative age dating of rocks due to the robust, well-mixed homogeneous nature of strontium isotopes in marine systems. Unlike other isotopic systems with limited residence times, strontium has a relatively long residence time and is uniformly mixed in the global ocean on short time-scales (~1000 years), within subsidiary marginal basins and seas, and even within some estuaries (McArthur, 1998). Strontium isotopes are also not subject to significant fractionation processes. Today, the global value of $^{87}Sr/^{86}Sr$ is .709175 (McArthur, 1998) and is observed in seawater around the world. The isotopic variation of $^{87}Sr/^{86}Sr$ through time relates to three primary controls: 1) weathering and transport of strontium from continental crust (average river flux is 0.712); 2) hydrothermal interaction within basalts at mid-ocean ridges (average MOR flux is 0.704); and 3) discharge of pore fluids from the seafloor after recrystallization of carbonate sediments (average seafloor flux is 0.708) (Tucker & Wright, 1990). Riverine input of strontium exceeds hydrothermal input by a factor of three and recrystallization flux is 1/3 less than hydrothermal input – therefore the primary driving forces for strontium isotopes in the ocean are dominated by terrestrial supply of sediments with some influence from mid-ocean ridge activity and minor recrystallization fluxes.

Toward a new integrated, high-resolution $^{87}Sr/^{86}Sr$ curve for the Upper Ordovician.

As mentioned the long-term decrease in strontium values during the Ordovician, especially in the Upper Ordovician (Ashbyan-Chatfieldian) has been equated with the large-scale transgression of the Tippecanoe Megasequence. This decline of $^{87}Sr/^{86}Sr$ is thought to reflect the decline of terrestrial sediment sources (high $^{87}Sr/^{86}Sr$ values) due to flooding of sediment source areas and/or a relative increase in the rate of hydrothermal delivery from mid-ocean ridges (low $^{87}Sr/^{86}Sr$ values). This might be expected as a result of increased rates of seafloor spreading during this time (Qing et al, 1998; Denison et al., 1998; Shields et al, 2003). In an effort to further refine the $^{87}Sr/^{86}Sr$ isotopic curve, data points from over 70 different samples from lower Upper Ordovician strata around the GACB (Oklahoma, Ohio, Tennessee, Ontario, New York, Kentucky, Michigan, and Missouri) have been placed in relative stratigraphic order using biostratigraphic placements where possible and approximations based on proximity to other important stratigraphic marker intervals.

As shown in **figure 18**, the combined data sets clearly show the long-term decline in

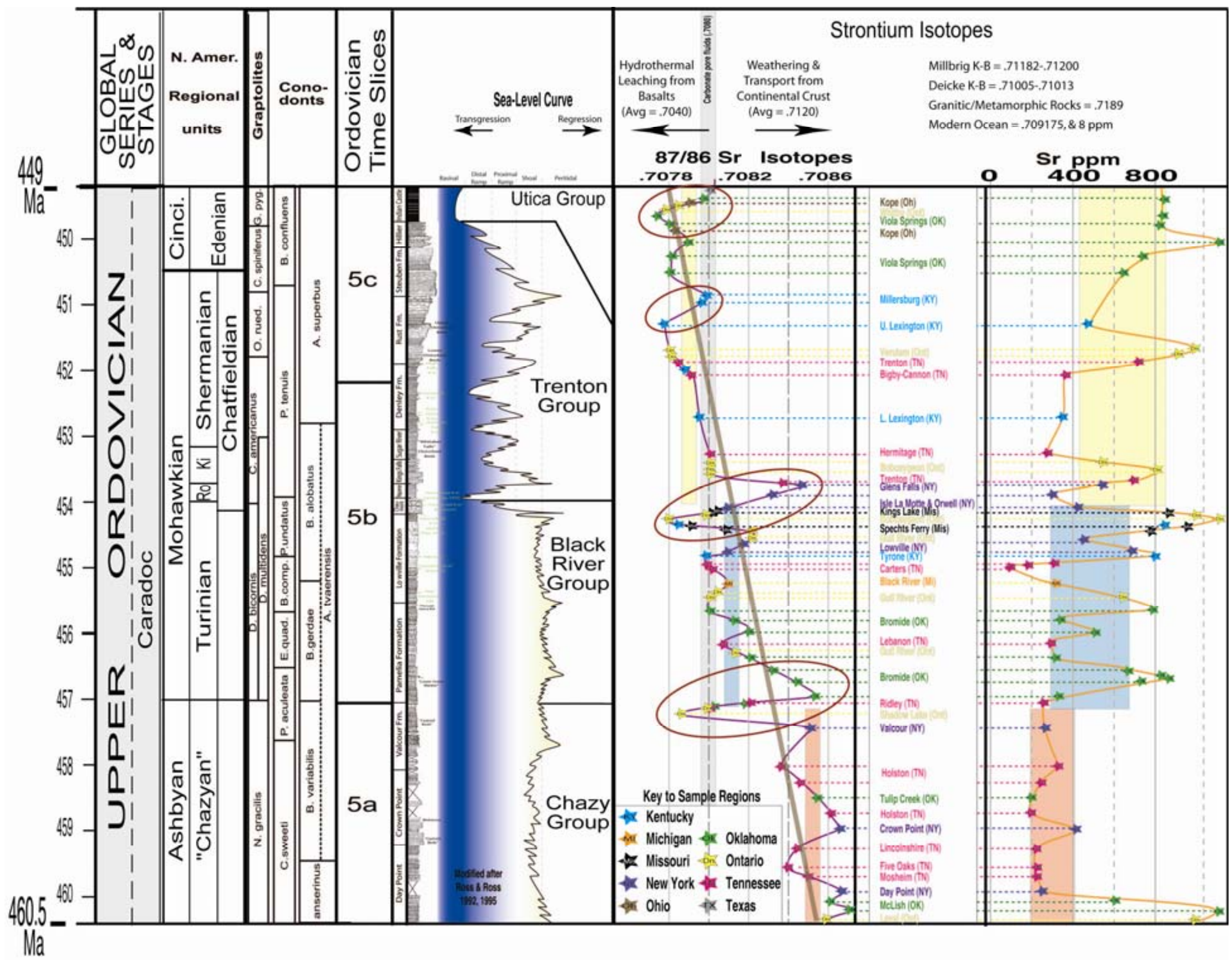


Figure 18: Integrated Strontium Isotope data for the Earliest Late Ordovician (the CBRT interval). Data are taken from Denison et al., 1998, Qing et al., 1998, and Shields et al., 2003, and integrated based on biostratigraphy, and stratigraphic position relative to other key marker units.

strontium isotope compositions from the base of the Chazy Group through deposition of the latest Trenton Group before values begin to recover again in latest Ordovician strata (not shown). This long-term decline clearly reflects a long-term change in the equilibrium between primary controls (i.e. less continental-sourced strontium – more marine-sourced strontium). Interestingly, when considered at the group scale, these data show that each of the major groups (i.e. Chazy, Black River and Trenton) each retain a characteristic range of values. For instance, the majority of Chazy equivalents have a range of values for $^{87}Sr/^{86}Sr$ in the range of .7084 to .70865 – with an average value of about .70855 for the group. For the Black River the values typically range between .7078 and .7085 with an average value for the group near .7081. The Trenton as a unit has values ranging from a low of around .7077 to .7080, with an average close to .7079.

Recognition of distinct $^{87}Sr/^{86}Sr$ isotopic excursions for the Upper Ordovician.

In addition to the long-term decline in $^{87}Sr/^{86}Sr$ values from Chazy through Trenton, if a best fit line is drawn through the data points to represent the equilibrium value change, a number of statistically significant excursions become noticeable. Discrete intervals within this ~ten million year period show that the $^{87}Sr/^{86}Sr$ equilibrium curve is interrupted by several relatively short-duration excursions of about one million years or less. These excursions appear to be unrelated to diagenesis and may represent previously unrecognized events or perturbations in the global $^{87}Sr/^{86}Sr$ curve. McArthur (1998) indicated that small but abrupt changes in the marine $^{87}Sr/^{86}Sr$ values, such as these, might occur in response to sudden, rapid changes in climate, and or rapid tectonic pulses. Hence it is possible that these excursions represent significant short-term climatic and/or tectonic events that punctuated the longer-term decline in marine strontium isotopic values.

Models for producing short-term positive $^{87}Sr/^{86}Sr$ excursions.

Although more work needs to be done to provide a higher resolution and detailed analysis of these excursions, given that the long-term decline in $^{87}Sr/^{86}Sr$ values is attributed to global transgression during the Creek Phase of the Tippecanoe Megasequence– it is likely that short-term excursions with increases in $^{87}Sr/^{86}Sr$ values would correlate with phases of sea-level drop and or rapid climate change producing increased runoff. Alternatively positive changes could also signify phases of tectonic uplift where continental and metamorphic rocks were freshly uplifted and exposed (Tucker & Wright; 1990; reported average $^{87}Sr/^{86}Sr$ values of .7189-.7198 for metamorphic and feldspar-rich granitic crust). A final possibility is that positive excursions could result from delivery of significant wind-blown sediment in the form of weathered continental rock dust or felsic to intermediate volcanic ash (Carey, 2006; reported $^{87}Sr/^{86}Sr$ values for the Deicke and Millbrig K-bentonites .71005-.71013 & .71182-.71200 respectively). As suggested by Dott (2003), the presence of windblown quartz-rich sediments in Chazy equivalent strata suggest significant contribution of wind-blown strontium-rich materials at this time.

Given denudation rates of modern uplifted terranes, it is likely that short-term excursions could be produced in response to significant tectonic uplift – for example see discussion on modern $^{87}Sr/^{86}Sr$ isotopic curve and the development of the Himalaya (Edmond, 1992). Short-term positive excursions could also feasibly correspond to: 1) global sea-level lowstand events with durations about one million years resulting in increased surface exposure and weathering of continental and/or metamorphic rocks globally; 2) dramatic climatic change where increased quantities of continental/metamorphic derived sediments were delivered to the ocean via wind-blown dust (aridity events) or via increased runoff (from monsoonal precipitation or melting events); or 3) delivery of large volumes of felsic to intermediate volcanic ash within a short-time scale. The later scenarios are somewhat problematic on the basis of mass-balance in that a large quantity of dust, ash, or sediments is necessary to achieve significant shifts in the marine

$^{87}Sr/^{86}Sr$ isotopic composition. Dust, ash, and sediment loads would need to approach the total volume of sediment supplied by rivers to the global ocean today.

Nonetheless as reported in the literature the increase in the number and volume of eruptive events in the late Turinian to early Chatfieldian globally (Huff et al., 1992; Kolata et al., 1996) suggest these events may have been significant enough to produce a $^{87}Sr/^{86}Sr$ isotopic change that exceeded equilibrium values for a short-time. In addition, Israelson and Buchardt (1999) report $^{87}Sr/^{86}Sr$ isotopic values from rivers draining the western side of the modern Greenland Ice Sheet and report significant positive fluxes in strontium where values for the region range up to 0.71963 and exceed average riverine values of the world's rivers. Therefore melting of large volumes of glacial ice over wide regions, during sea-level rise could potentially dampen or overwhelm equilibrium $^{87}Sr/^{86}Sr$ values and effectively produce a short-term positive excursion even as sea-level is rising.

Lastly, although strontium supplied to the global ocean in wind-blown dust is much harder to quantify, a significant number of studies have used $^{87}Sr/^{86}Sr$ isotopic values to suggest the role and dominance of wind-blown dust in supplying sediment to coastal regions in arid environments and indeed to many islands. For example in Corsica, the Cape Verde Islands, and other islands in the sub-tropical Atlantic, low $^{87}Sr/^{86}Sr$ values from weathering of basalts have been buffered in soils and coastal sediments by high $^{87}Sr/^{86}Sr$ from wind-blown Saharan sources and result in significantly positive values (Rognon et al., 1996). In this case, and in others, the variable record of $^{87}Sr/^{86}Sr$ has been used to calibrate frequency of aridification/monsoonal oscillation events in Saharan Africa. It is not currently known to what degree wind-blown materials influence the $^{87}Sr/^{86}Sr$ values of the global ocean, but in the absence of substantial runoff during lowstand and/or periods of low seafloor spreading, aridification could potentially be the dominant source of continental sediment delivered to the ocean and could result in short-term positive $^{87}Sr/^{86}Sr$ excursions.

Models for producing short-term negative $^{87}Sr/^{86}Sr$ excursions.

In addition to short-term positive excursions in $^{87}Sr/^{86}Sr$ isotopic values for the CBRT, investigation of the combined data set also produces several short-term excursions characterized by lower $^{87}Sr/^{86}Sr$ ratios. As indicated previously, the lowering of $^{87}Sr/^{86}Sr$ values is favored when strontium sources become dominated by increased volumes of basalt influenced by hydrothermal leaching. Most commonly this occurs at mid-ocean ridges on the sea-floor, but it may also occur in areas where major basalts become significant sources of sediment, or in areas where basaltic tuffs supply a significant volume of sediment. Another potential mechanism mentioned previously could include massive discharge of pore-fluids from diagenetically altered carbonate deposits. However, this scenario is discounted on the basis that the flux of pore-fluids is typically very small and $^{87}Sr/^{86}Sr$ values of discharged pore-fluids average only about .7080 – which is not low enough to explain the lowered $^{87}Sr/^{86}Sr$ excursions in the CBRT interval.

Recognition of Paired $^{87}Sr/^{86}Sr$ excursions.

Given the $^{87}Sr/^{86}Sr$ strontium equilibrium trend shown in **figure 18** for the CBRT interval although tentative, it appears that there are four dominant paired excursions that can be recognized as significant. These paired excursions are recognized as a fairly sharp lowering of $^{87}Sr/^{86}Sr$ ratios followed by a rapid-sharp positive excursion within an interval of 1-2 million years before settling into near equilibrium trajectories. The four paired excursions thus correspond with the Chazy-Black River transition, the Black River – Trenton transition, the lower-upper Rust transition (Upper Lexington to Point Pleasant of KY-OH), and the lower Indian Castle (Kope of KY-OH). The occurrence of negative followed by positive excursions is intriguing as in all cases there is an accompanying pronounced lithologic change that is the basis for the recognition of the lithologic boundary itself.

Ashbyan-Turinian Boundary $^{87}Sr/^{86}Sr$ excursion.

The first excursion is recognized from data points taken from strata in New York, Ontario, Oklahoma, and Tennessee. During the latest Ashbyan medial *P. aculeata* conodont chronozone, $^{87}Sr/^{86}Sr$ ratios drop from near the average Ashbyan values of .7085 to values characteristic of the Trenton at just over .7078. These low values are recorded from several conodont apatite samples from the Shadow Lake and Gull River Formations of Ontario (Qing et al., 1998 data). As recognized from samples in the Ridley Formation of Tennessee and from the Bromide of Oklahoma (Denison et al. 1998 data from carbonate sediments) $^{87}Sr/^{86}Sr$ ratios then shift rapidly upward in the early Turinian to values characteristic of the Chazy (.70858) before dropping again to characteristic Turinian values. This excursion records the largest $^{87}Sr/^{86}Sr$ ratios for the entire Black River interval. The low values are not reached again until the lower Trenton during the second excursion.

Thus, the first (declining) phase of the paired excursion coincides with facies deepening associated with flooding of the GACB including portions of the Canadian Shield where the conodonts were sampled. The second (positive) phase of the paired excursion coincides with facies shallowing leading into deposition of the Black River Group. Thus in this case, the excursion could potentially be tied to a short-term sea-level oscillation. Alternatively, if the positive values are considered a hold-over from Chazy values and are considered part of the normal equilibrium – then the declining excursion that predates it could be considered independently and therefore related to a significant pulse in seafloor hydrothermal activity perhaps associated with the Blountian Tectophase.

Black River - Trenton Boundary $^{87}Sr/^{86}Sr$ excursion.

The second paired excursion set begins in the top of the Black River Group and ranges into the lower Trenton and thus appears to be nearly coeval with portions of the Spechts Ferry and Guttenberg Isotopic Carbon excursions discussed earlier. The paired excursion is recorded from strata in Ontario (Gull River – Bobcaygeon), Missouri (Spechts Ferry – Kings Lake), Kentucky ("Bobcaygeon"-Curdsville), New York (Isle La Motte-Orwell-Glens Falls), and Tennessee ("lower Trenton"). In this case characteristic Black River $^{87}Sr/^{86}Sr$ ratios in brachiopods from the shale-dominated facies of the Spechts Ferry drop rapidly. The lowest values (~.7075) are recorded in the lower Bobcaygeon of Ontario (data from Shields et al., 2003). Subsequently the largest $^{87}Sr/^{86}Sr$ ratios are recorded in the lower Trenton of Tennessee (Hermitage) and the lower Trenton of New York (Glens Falls Limestone) where values reach .7085 (Denison et al., 1998). $^{87}Sr/^{86}Sr$ ratios values rapidly return to characteristic Trenton strontium values ranging between .7078 and .7079.

Thus defined, the second paired excursion shows its declining phase in the latest Turinian and is coincident with the eruptions responsible for deposition of the Millbrig K-bentonite within the overall shallowing phase at the end of the Black River. The ensuing rapid positive shift reaches a maximum near the Kirkfieldian boundary which coincides with shallowing of facies immediately following earliest Trenton deepening. Nonetheless – the rapid rise of this excursion initiates during rapid transgression. In contrast to the first $^{87}Sr/^{86}Sr$ excursion of the CBRT – the second $^{87}Sr/^{86}Sr$ excursion appears out of phase with sea-level change. In this scenario the declining phase, suggestive of increased rates of sea floor hydrothermal activity, is coincident with the large-scale volcanic eruptions of the Millbrig interval. Although the contribution of the Millbrig and related ashes would likely elevate the $^{87}Sr/^{86}Sr$ ratios – values still drop to very low levels suggesting major sea-floor spreading or hydrothermal activity initiated coincident with these volcanic eruptions. As suggested by Karabinos and colleagues (1998), this may have been a time of slab break-off and subduction reversal off eastern Laurentia, which may have allowed for extremely high sea floor hydrothermal activity leading to the Vermontian tectophase of the Taconic Orogeny.

Subsequently, the positive phase of the $^{87}Sr/^{86}Sr$ excursion initiates its shift during a deepening phase before maxing out in a shallowing interval. This suggests that the supply of strontium shifted to terrestrial sources either due to rapid tectonic uplift and initiation of erosion or climatic change where a large supply of sediment globally was delivered to the oceans. As mentioned previously these could have been achieved by: increased runoff (induced by high precipitation), increased wind-blown dust (aridification), or rapid melting of sediment-laden glacial systems that would not necessarily be tied to sea-level change.

Medial *O. ruedemanni* zone $^{87}Sr/^{86}Sr$ excursion.

The third excursion is recorded by strata from the Upper Lexington Group (Millersburg Limestone; Denison et al., 1998), the upper Verulam of Ontario, and the Bigby-Cannon Limestone of Tennessee. The paired excursion is less pronounced compared to both earlier excursions and is recognized relative to typical Trenton $^{87}Sr/^{86}Sr$ values. The initial, depressed phase of this excursion shows declining $^{87}Sr/^{86}Sr$ ratios in Tennessee, Kentucky and Ontario to the lowest values of the Trenton (just less than .7078) in the central Jessamine Dome before rebounding in the second phase of the excursion to about .7080. In this scenario, the depressed phase again appears to coincide with an interval of overall deepening of facies before shallowing in the second phase of the excursion. This pattern is thus in phase with sea-level change and may simply reflect the expected response associated with deepening (shutting down the supply of terrestrial sediments with high $^{87}Sr/^{86}Sr$ values) and subsequent shallowing (increasing the supply of terrestrial sediments).

G. pygmaeus zone $^{87}Sr/^{86}Sr$ excursion.

The final excursion in the CBRT interval of this study occurs immediately following deposition of the Trenton and its equivalents and is defined in strata from Ontario (lower Whitby), Ohio (Kope Formation), and Oklahoma (upper Viola Springs). This $^{87}Sr/^{86}Sr$ excursion begins its decline from the average Trenton Value of around .7079 and drops to the lowest value of the entire CBRT at about .7077 and then gradually climbs above the Trenton average to values above .7080 in the Kope Formation, and the Whitby interval. In higher strata, as reported by Denison and colleagues (1998), values high in the Viola Springs and other equivalent formations show a decrease again to near .7077 to .7078 suggesting that values return to near the equilibrium trend before recovering in later Ordovician strata. Thus this particular excursion initiates its first, declining phase in the Trenton during a relative deepening phase. Its positive increasing component appears to coincide with highstand deposition in the Kope. Therefore as with the first and third excursions, the fourth $^{87}Sr/^{86}Sr$ isotope excursion appears to correlate well with sea-level change and the delivery of siliciclastic sediments that are clearly being contributed to the GACB from the Taconic Highlands after the demise of the GACB.

Implications of $^{87}Sr/^{86}Sr$ excursions.

As discussed, the recognition of these $^{87}Sr/^{86}Sr$ isotopic excursions is only tentative and awaits further analysis and testing. Nonetheless, considered within a detailed high-resolution chronostratigraphic framework, the potential for further exploration of the nature and timing of these excursions is substantial. Moreover, if calibrated to an independent chronostratigraphic framework, strontium isotope stratigraphy could serve as an additional method for correlating stratigraphic sections across even greater regions owing to the excellent homogenization of $^{87}Sr/^{86}Sr$ values in the global ocean and marginal seas. Moreover, this method may also provide an excellent tool for investigating the relative response of sea-level, lithologic change, and the potential impacts of tectonics and climate change within geologically complex regions including foreland basin settings. Thus this system might provide additional evidence that could help unite structural/tectonic evidence available from the hinterland with stratigraphic evidence from the foreland basins themselves.

Within the GACB, the development of the composited excursion herein is possible based on the assumption that the $^{87}Sr/^{86}Sr$ values across the region behave as predicted by previous authors – that is they are well-mixed and reflect global trends. Although future work should focus on the documentation and correlation of the curve regionally, preliminary work discussed above points to several plausible and important chemo-stratigraphic events that likely were tied to climatic and or major tectonic pulses. It is especially important to mention the occurrence of the first two paired excursions and their coincidence with important lithostratigraphic and tectono-stratigraphic intervals in the CBRT interval, in particular the activation of the Blountian and Vermontian tectophases respectively.

Nd-Isotopes:

General Introduction to Neodymium Isotopes in the Ordovician

Neodymium isotopic analyses of ancient rocks has become a robust method for not only establishing reliable estimates of ancient seawater composition at the time of deposition (Holmden et al., 1998), but has also been increasingly used to establish provenance (Gleason et al., 1994, Fanton et al., 2002), and for sequence recognition and correlation (Ehrenberg et al., 2000; Fanton & Holmden, 2001; Fanton et al., 2002; Fanton & Holmden, 2007). As rocks age, ^{147}Sm undergoes radioactive decay to the stable ^{143}Nd isotope, thus older rocks contain significantly higher concentrations of the decay product. In contrast ^{144}Nd is stable and not susceptible to change from values found in original source rocks. Therefore as rocks age their ε_{Nd} values become more negative. ε_{Nd} values are determined relative to a chondrite standard (DePaolo & Wasserburg, 1976 a,b) and are reported in parts per 10,000 deviation from the standard. Hence, the ε_{Nd} of continental crust is age dependent and variations in the concentration of ^{143}Nd and ^{144}Nd isotopes reflect the age of the rocks making up the continental crust. Thus when continental rocks of different ages weather and their weathering products are transferred to the ocean, the ε_{Nd} values of seawater in the region near the source of those sediments takes on the ε_{Nd} values inherited from the source terrane. Moreover, as reported by Piepgras and Wasserburg (1980), neodymium isotopes have a very short residence time in modern seawater (less than 2 in surface waters to between 100 and 300 years in deeper waters; Jacobson & Holmden, 2006) and a relatively short mixing time (approx. 1000 years). Therefore neodymium isotopic composition is rarely homogenized in the global ocean and especially within marginal seas – unlike other isotopes that have much longer residence times. The neodymium composition of the global ocean and marginal seas is thus strongly influenced by continental weathering and sediment transport (as dust or fluvial-transported sediment) to the oceans especially during times of lowered sea-level when exposed source terranes are variable.

Wright and colleagues (1994, 2002) investigated the behavior of neodymium isotopic variation ($^{143}Nd/^{144}Nd$) for Ordovician seas globally. In their study, ε_{Nd} isotopic values were determined using samples of conodonts taken from epicontinental sea deposits of major cratons and suspected microplate terranes. Conodonts are thought to record and preserve the ambient isotopic composition of seawater within which they grew. As a result of their investigation, it was shown that during the Ordovician significant differences existed in the ε_{Nd} composition of the Laurentian epeiric sea when compared to other major structural provinces (**figure 19**). Specifically ε_{Nd} values indicated that Early Ordovician water masses associated with Laurentia were strongly negative (-28 to -18) when compared to most other cratons – thus suggesting this region was sourced by significantly older continental crust – than were other contemporaneous craton regions. These negative ε_{Nd} values rebounded to -17 to -11 at the end of the Llanvirn (pre-Knox unconformity ~470 mya) and became more similar to values observed on other cratons including Baltica, South China, and Peri-Gondwana terranes. This suggests that the sea-level highstand before the close of the Sauk Supersequence produced the conditions necessary for the neodymium isotopic composition of marginal seas to become much more similar to those of other cratons and that younger sediment source terranes became substantially more important. However, the Wright et al. (2002) data show that immediately following the pre-Knox highstand ε_{Nd} values in Laurentia again declined sharply in the base of the Caradoc (~460 mya) to values of -21 to -18 during the 5a time-slice (of Webby et al., 2004), suggesting isolation of Laurentia once again from global values and an increase in the age of sediment sources owing to lowered sea-level. Subsequently, values again began to rise throughout the rest of the Ordovician to ε_{Nd}

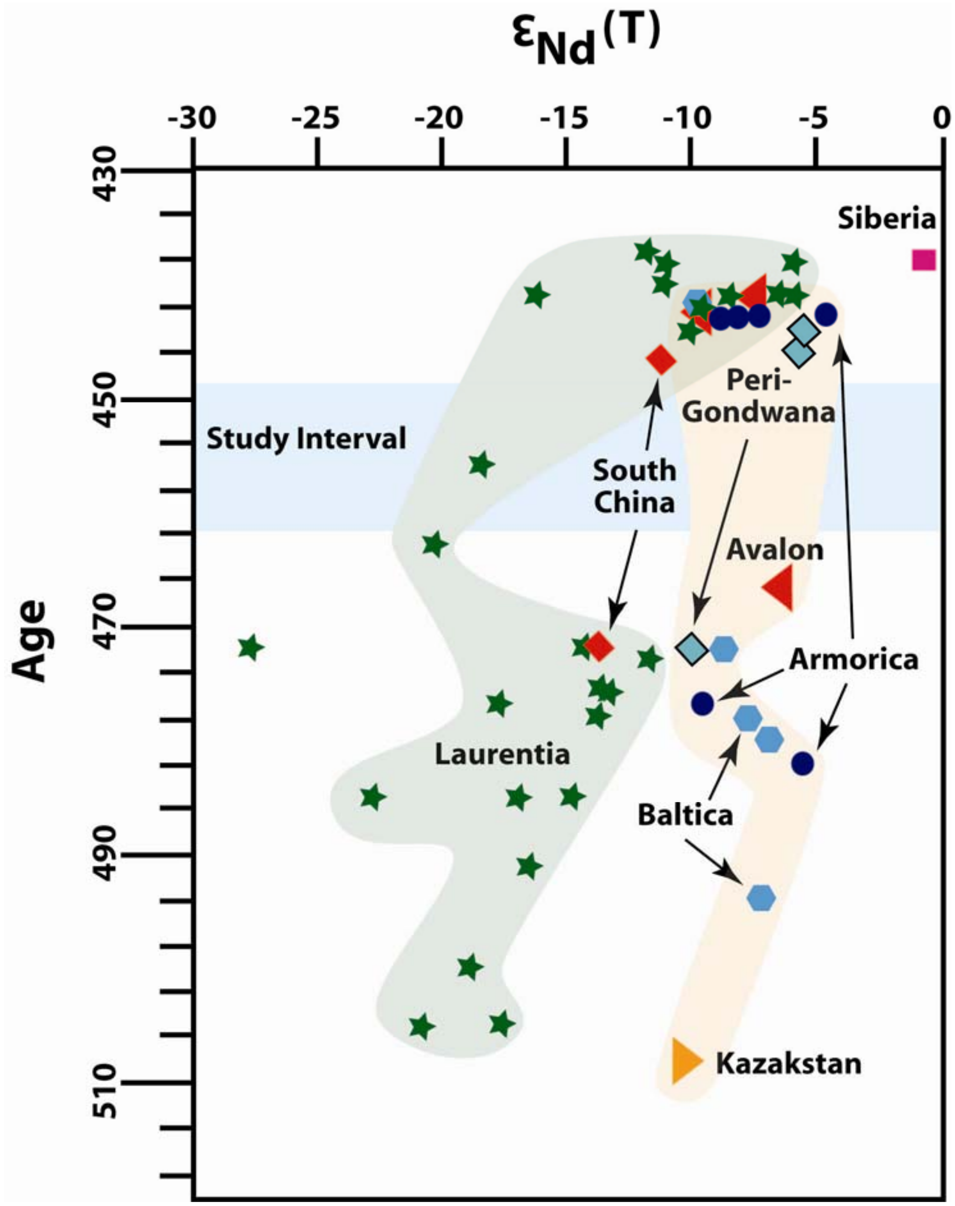


Figure 19: ε_{Nd} values for the Ordovician to earliest Silurian. Shown in the shaded green region are the ε_{Nd} values for the Laurentian platform region, shown in pale tan are the ε_{Nd} values for peri-Laurentian cratons and micro-plates. Shaded in light blue is the approximate interval investigated in this study. As shown, peri-Laurentian ε_{Nd} values typically range between -10 and -5 and Laurentian ε_{Nd} values show evidence for decoupling from more regional values. Figure modified from Wright et al., 2002

values of between -12 to – 5 in the latest Ordovician reflecting a significant shift in the age of sediment source terranes and higher sea-levels. Fanton and colleagues (2002) recognized a particularly sharp rise in ε_{Nd} values going into the 5c time-slice (of Webby et al., 2004) reflecting again a major change in age of sediment source terranes. In the latest Ordovician, ε_{Nd} values are thus nearly identical to all major cratons and suggest that oceanic water-masses were again highly uniform in the ε_{Nd} values.

Neodymium Isotopic Variation Across the GACB

As suggested by Holmden and colleagues (1998), there is substantial evidence for decoupling of the Laurentian ε_{Nd} isotopic values from that of the global ocean suggesting the strong influence of local sedimentary provenance of the GACB region over global oceanic isotopic values (see **figure 19**). Holmden and colleagues (1998) looked at the narrow time slice between the Deicke and Millbrig K-bentonites and recognized a regional variation across the GACB from the mid-continent region to the Taconic Foreland Basin. Conodont apatites showed a gradient of strongly negative ε_{Nd} values from the vicinity of the Transcontinental Arch with subsequent increase in ε_{Nd} toward the Taconic Foreland where ratios reached maximum values approximating those of the Iapetus Ocean (ranging to about -3). These authors suggested that the negative values from the Transcontinental Arch region (-17 to -19), their "Midcontinent aquafacies," were derived from weathering of 2.1 to 2.7 billion year old rocks from the Transcontinental Arch and Superior Province portion of the Canadian Shield (Holmden et al., 1998). Values from the "Taconic aquafacies" are less negative (-6 to -9), and are thought to have been derived from the younger (1.7 to 1.9 billion year old) Grenville Province sources along the eastern Canadian Shield and from within the Taconic Orogen itself. This assessment is supported by Bock and colleagues (1998) who investigated ε_{Nd} values from the Austin Glen Formation deposited on the continental slope and rise east of the New York promontory during the CBR. These data suggest the Grenville Province was the dominant source terrane for sediments on the Laurentian slope leading into the Taconic Orogen. The former authors also recognized a "Southern Aquafacies" that records ε_{Nd} values more typical of the Iapetus Ocean (-3 to -5) which suggests Laurentian or Taconic-derived sediments were not as important in contributing neodymium to water-masses in these regions and water masses here were more closely linked with open marine waters.

Neodymium Isotopic Variation with Low-Order Sea-Level Change

As noted previously, data from Wright and colleagues (2002), when considered in tandem with large-scale patterns of sea-level during the Ordovician, suggest the potential for changes in ε_{Nd} values to reflect major changes in sea-level. Ehrenberg and colleagues (2000) suggested that neodymium isotopic profiling in carbonate platform strata for the Pennsylvanian to lower Permian could be used as a method for correlating siliciclastic provenance with sequence stratigraphy. In this study, it was documented that significant changes in $^{143}Nd/^{144}Nd$ ratios could be equated to major 2nd order sea-level changes owing to drowning of local sediment sources (Caledonian basement). During sea-level rise $^{143}Nd/^{144}Nd$ ratios increased and sediments were supplied by open-marine siliciclastic transport from a wider region and were more reflective of average oceanic values. In contrast during sea-level fall, $^{143}Nd/^{144}Nd$ ratios were shown to decline and reflected influx of relatively "old" locally-derived basement material.

Thus, the trends in ε_{Nd} values (standardized $^{143}Nd/^{144}Nd$ ratios) presented by Wright and colleagues (2002) might be considered as reflective of the variation expected by termination of the Sauk Megasequence and initiation of the Tippecanoe Megasequence (see **figure 19**). Specifically, the increase in ε_{Nd} values into the medial Ordovician, Darriwilian Global Stage, (Laurentian ε_{Nd} values approaching average global values) coincides with maximum sea-levels prior to the Knox Unconformity. This trend suggests that Laurentian-derived signatures became less substantial as the Transcontinental Arch and Canadian Shield became flooded during widespread development of the Laurentian carbonate platform. During the Knox, and subsequently during the Sandbian and Katian Global Stages (Upper Ordovician), sea levels initially dropped and they began to recover. Coincident with sea-level, ε_{Nd} values once again became strongly negative and again show a rapid upward increase in ε_{Nd} values into the Upper Ordovician. The shift to more negative values (and increased departure from average global values) suggests that the sediment supply of Laurentia became dominated once again by sediments derived from the Transcontinental Arch and portions of the Canadian Shield as they became exposed during the Knox unconformity. Subsequently during the Tippecanoe (Creek Phase) transgression, Laurentian cratonic ε_{Nd} became swamped by significantly younger sediments derived from the newly uplifted Taconic Highlands.

Neodymium Isotopic Variation with High-Order Sea-Level Change

Although Ehrenberg and colleagues (2000) were unable to identify any higher-order signals in their data set, i.e. they were unable to ascertain variations in neodymium that would enable them to recognize shorter-frequency sea-level oscillations as previously established by sequence stratigraphy), Fanton and colleagues (2002) and Fanton (2004) reported for the first time the correlation between ε_{Nd} fluctuations and high-order sea-level change. In these studies, it was shown that ε_{Nd} values for Chatfieldian through Richmondian strata of Iowa not only show the long-term trend to less negative ε_{Nd} values through the interval, but also show short-term fluctuations in ε_{Nd} values. Fanton and Holmden (2007) identified 6 excursions (I1 through I6) ranging from the Decorah up through the Dunleith into the Wise Lake Formation of Edenian age (**figure 20**). Excursions I1 through I6 correspond to the sea-level fluctuations recognized from the upper Mississippi Valley region by Witzke and Bunker (1996) and thus were interpreted to

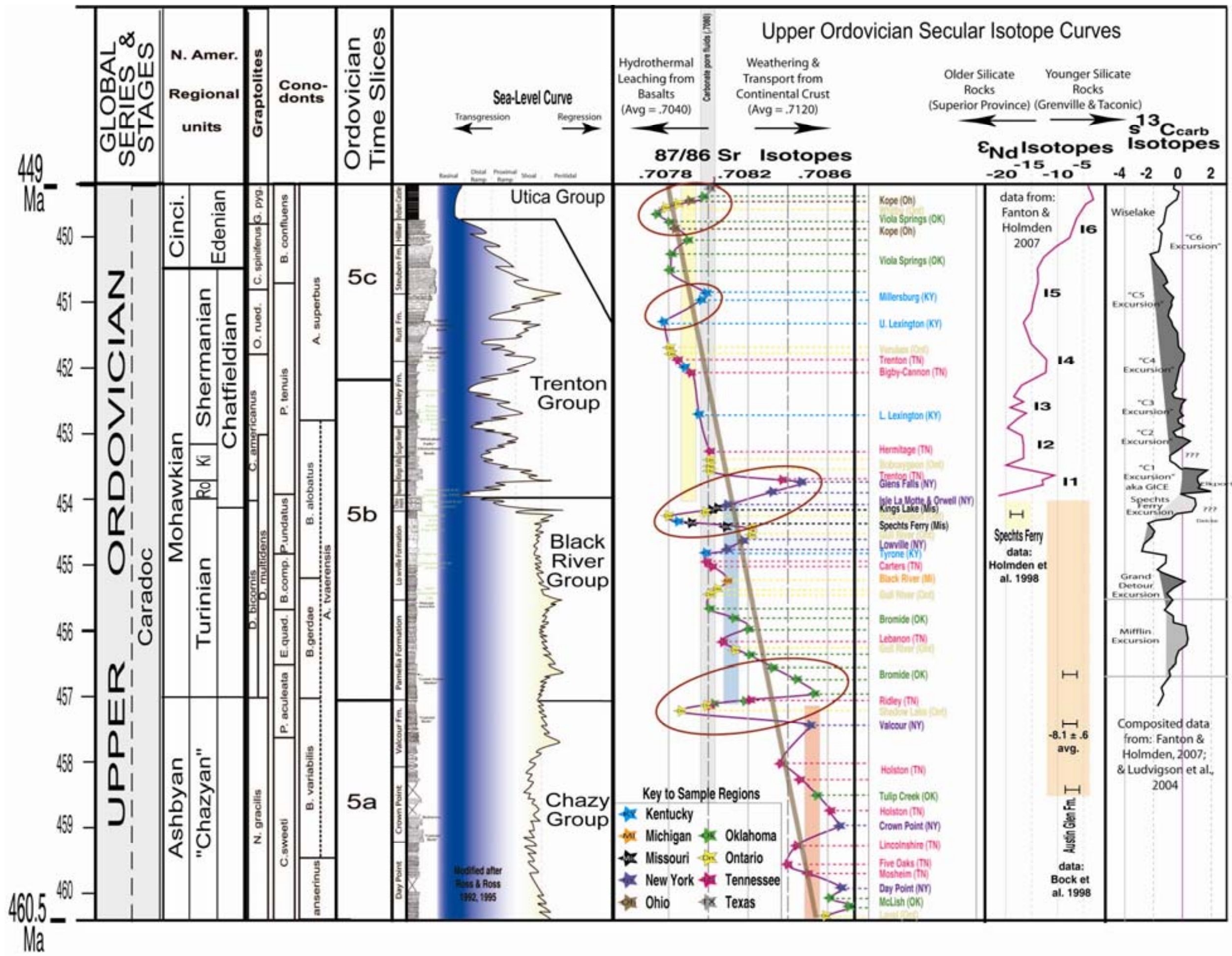


Figure 20: Upper Ordovician Secular Isotope Curves for eastern Laurentia. ε_{Nd} values are shown as are carbon isotopic excursions as recognized by Fanton & Holmden (2007). Data shows the range of values for Iowa as well as slope deposits for eastern North America during deposition of the Austin Glen Formation as reported by Bock et al., 1998.

represent sea-level oscillation at higher-order scales as recognized by depositional sequences.

The first three excursions (I1 to I3) show relatively little net increase in ε_{Nd} values overall, but fluctuate between -20 to -13, while the second three excursions (I4 to I6) show a net positive increase to less negative ε_{Nd} values, with a maximum at ~-5 during deposition of the middle Wise Lake Formation. The last excursion (previously recognized as I-4 in Fanton et al., 2002), coincides with a nearly identical and simultaneous excursion from southeastern Saskatchewan – a region which had until this time exhibited strongly negative ε_{Nd} values. From these data it is suggested that early Chatfieldian sediments from the Iowa region were dominated by Transcontinental Arch/Canadian Shield-derived signatures during periods of sea-level lowering (via evacuation of drowned estuaries). In contrast during periods of transgression, sediment supply from the craton became diminished due to sequestration and flooding of source areas. Simultaneously owing to the recent activation and uplift of the Taconic Highlands and exposure of Grenville and Grenville-derived sedimentary rocks, sediment signatures thus became dominated by the influx of sediment from the orogen. In later excursions subsequent to I3, the supply of neodymium from the Transcontinental Arch and the older provinces of the Canadian Shield are significantly reduced while neodymium sourced from the Taconic Orogen become dominant. This signature becomes so important by the Edenian that even sedimentation on the northwestern side of the Transcontinental Arch, in the vicinity of the Williston Basin in southern Saskatchewan, Canada, becomes influenced by Taconic-derived neodymium and or other average global sources.

Implications of Neodymium Isotopic Variation for Taconic Tectonism and Sealevel

Based on data presented by Fanton and Holmden, (2007), the I1 excursion has the greatest amplitude in ε_{Nd} values (-21 to -11) of all Chatfieldian excursions. It is matched only by the I6 excursion with ε_{Nd} values ranging from -15 to -6. The third largest amplitude excursion (I4, ε_{Nd} values ranging from -18 to -13) occurs in the middle of the Dunleith Formation and is coincident with a significant increase in both total organic carbon and $\delta^{13}C$ values of carbonates (C4 Excursion; Fanton & Holmden, 2007). Excursions I2, I3, and I5 are of significantly lower amplitudes with variation on the order of 2 to 3 epsilons.

The relative scales of these six excursions are thought to reflect the magnitude of sea-level rise for each individual interval based on lithologic and biologic indicators and or significant pulses in tectonism. Thus the appearance of the I1 excursion (which is synchronous with the appearance of the GICE, the tentatively recognized Turinian-Chatfieldian strontium excursion, and the initiation of Trenton-style deposition) signals the first major flooding of the entire GACB platform during the Rocklandian and signals the onset of the Vermontian Tectophase of the Taconic Orogeny. Subsequent excursions I2 and I3 are diminished relative to the first, and represent relatively small sea-level oscillations. Excursion I4 appears to coincide with a major flooding surface in the Rivoli Member of the Dunleith and is approximately coincident with the deposition of the lower Flat Creek Shale in the Taconic foredeep of eastern New York.

Thereafter, excursions I5 (diminished amplitude) and I6 show a long-drawn out increase in ε_{Nd} values from ~-17 to -5 during the latest Chatfieldian through the medial Edenian before declining again to more negative values. This long term increase in ε_{Nd} values coincides again with the minor sea-level fluctuations observed in the upper Dunleith (upper Trenton) followed by the large-scale transgressive phase leading into the Edenian. This excursion is thus coincident with deposition of the Point Pleasant to Kope interval of the Cincinnati Arch and the Steuben to Indian Castle Shale interval of New York. This final event represents the end of Trenton-style deposition and the eastern portion of the GACB region is thereafter inundated by siliciclastic influenced deposition as the Taconic Foreland begins to transition to flysch phase deposition especially in southern Virginia. There is potential therefore that the magnitude of change in ε_{Nd} values in this case represent another significant pulse in Taconic Orogenesis resulting in geomorphic changes in the structure of the foreland basin and orogenic hinterland that allowed an ever greater volume of sediment to be delivered to the Laurentian epicontinental sea.

SUMMARY AND IMPLICATIONS OF TIME-RESTRICTED FACIES FOR TECTONICS, SEA-LEVEL, AND CLIMATE CHANGE OF THE GACB.

As discussed herein the chronology of events in the Chazy, Black River and Trenton groups is now enhanced by recognition of associations of at least five different facies and nearly two dozen chemostratigraphic events. These collectively provide a substantially more detailed record of Late Ordovician environmental change than previously reported. With a much improved chronostratigraphic framework, the linkages between widespread lithologic change, changes in sedimentary provenance, sea-water chemistry, sea-level change, and associated pulses in tectonic activity are more strongly linked. The time-restricted facies documented herein support a "layer-cake renaissance" approach for use in foreland basin studies. Interestingly many of these events increasingly point to a number of allocyclic processes that influenced deposition across much of the GACB, while a number of key tectonic events are also evident.

Aside from documentation, for the first time, the nature of these time-restricted facies and their potential for use in correlation studies, this study suggests that many of these facies are excellent barometers for identification of unique climate and tectonic events, many of which have been poorly documented until now. Therefore an important contribution of this discussion involves the degree to which different tectophases of the Taconic Orogeny influenced environmental change in eastern Laurentia in regard to sea-level, water mass circulation patterns, possible climatic shifts, and at what relative times did these changes occur. The following outlines the most important and salient contributions of this study.

a) Recognition of four prominent distinctive lithologic facies associations that can be used for local to regional correlation. Moreover because they appear to be developed during relatively narrow windows of time when particular environmental conditions were developed, they also provide insights into sea-level, climate, oceanographic circulation, tectonics, etc. The five facies include:

- A range of dark blue-gray to black, nodular to slightly bedded cherts, fossil infilling cherts (& silicification), and or chalky white cherts. There are at least nine chert-rich intervals documented. Chert formation has generally been attributed to: 1) biologic production of siliceous skeletons, 2) upwelling of deep marine, silica-rich waters, 3) demineralization and reprecipitation of silica from volcanic ashes, 4) late diagenetic hydrothermal dissolution from other quartz-rich deposits and reprecipitation, 5) hot spring precipitation, and 6) recrystallization of silica-rich wind-blown dust.
- Siliciclastic events (SE's) that were intermittently deposited during periods of time when carbonate production dominated. Hence SE's are anomalous within the GACB region and have important implications. There are six SE's recognized prior to the major SE that signals the Vermontian Tectophase of the orogeny. Typically they are recognized by large contributions of quartz sand and/or clay/mud deposited in association with variable amounts of limestone, dolomitic limestone, and dolostone. Many have characteristic buff to greenish colorations owing to glauconitic-chlorite-rich compositions. Most SE's are developed during periods of sea-level fall (see Chapter 7) or possibly lowstand conditions thus delivery of siliciclastic materials is either late-stage progradation or possibly wind-blown as appears to be recorded in some SE's.
- Calcification events (CE's), which are composed of stromatolite and ooid-rich intervals and even show occasional development of evaporitic textures and structures indicative of evaporative conditions. CE's also show higher numbers of intra- and extraclastic limestone conglomerate and breccia beds. There are two pronounced CE's in the CBR

interval and likely indicate periods of major climate change (aridification) and hyper-calcification episodes. They may also record the loss, temporarily, of diverse marine communities and/or intense bacterial "bloom" events.

- Syn-sedimentary deformed intervals, interpreted often as "seismites." These deformed intervals are most prevalent in laminated to rhythmically-bedded calcisiltite to fine-grained grainstone and shale ("pin-stripe") facies, which were developed during regressive conditions and typically predate transgressive (subsidence) phases. These facies, due to their abundant shaly interbeds, were apparently more likely to undergo liquefaction during seismic shaking events, which although present in the Sevier Basin area, were much more common across the GACB beginning in the early Chatfieldian. These facies are particularly tied to major episodes of faulting and topographic alteration of the GACB during pulses of tectonism.

b) Herein a number of previously published chemostratigraphic data sets (Carbon, Strontium, Neodymium) have been integrated and synthesized using the compendium of stratigraphic correlations documented herein. Especially important has been the construction of a much more refined strontium isotopic curve for the CBRT interval. As a result, a number of important implications have been discussed related to changes in sedimentary provenance, ocean circulation, tectonism, sea-level and climate.

- Carbon isotopes

    i. Overall trend from light to heavy $\delta^{13}C$ isotopic ratios during the late Ordovician with increasing quantities of ^{12}C relative to ^{13}C being moved out of oceanic reservoirs (i.e. biological fractionation) and stored elsewhere in sediments.

ii. The Late Ordovician trend is punctuated in the Turinian-Chatfieldian with at least ten positive excursions each separated by negative excursions where light carbon is re-introduced into oceanic reservoirs.

- In the Ordovician, strong positive (global) excursions are thought to result from substantial removal of ^{12}C from the surface ocean to: 1) the deep ocean during ocean stagnation and anoxia; or 2) the ocean floor during periods of rapid sedimentation and burial. Alternatively, light, biologically fractionated carbon can be temporarily stored in sea-ice ("green ice"). Green ice forms by under-plating of ice shelves as they grow in the vicinity of highly productive, nutrient-rich upwelling zones. Therefore they have the potential to trap significant amounts of phytoplankton and hold that carbon well into the next warming phase.
- Slight positive excursions, locally or over a small region in a restricted basin, can result from increased riverine flux of bicarbonate (HCO_3-) from weathering of silicate rocks and or dissolution of exposed carbonate rocks. These are likely the origin of many of the isotopic excursions in the Turinian.
- Negative shifts (the return of ^{12}C to the oceanic reservoir) can result from: 1) re-invigoration of ocean circulation and upwelling of deep, light C; 2) volcanic degassing; 3) weathering and transport of previously deposited organic-rich rocks; or 4) release of methane gas from the seafloor during warming or shallowing.

iii. The most prominent excursion is the prominent early Chatfieldian (Guttenberg Isotopic Carbon Excursion or GICE) event with $\delta^{13}C$ values reaching a maximum of ~4‰, although more typically ~ 2‰. It is thought to represent a global event (see Saltzman et al., 2003).

- ° The GICE shows lightest values near the Transcontinental Arch with heavier values toward the Sebree Trough and highest values in the Taconic Foreland Basin. This trend was interpreted to result from upwelling of nutrients, greater rates of biological fractionation, and increased organic carbon burial in the latter regions comparatively (Fanton & Holmden, 2007).
- ° The GICE initiates synchronously on GACB and in the Sebree Trough and the Taconic Foreland Basin, although heavy values persist through the entire Chatfieldian in the foreland basin. Likely the extended excursion in the foreland reflects composited excursions that are five distinct excursions (C1-C5) on the platform.
- ° Thus, evidently the foreland basin became decoupled from the GACB with respect to water mass circulation during the early Kirkfieldian when the GICE ends on the platforms (and in the Sebree Trough) to the west of the foreland. It remained decoupled until at least the end of the Chatfieldian.

iv. The GICE occurs during the end of the first transgression of the Chatfieldian (Rocklandian) and persists through the ensuing regression into the base of the next transgressive phase (earliest Kirkfieldian).

- ° Thus the GICE, globally, likely reflects a major episode of ocean stagnation during a major eustatic rise event and carbon isotope partitioning. By implication the earliest Chatfieldian (Rocklandian-Kirkfieldian) may record a short-term (~<1my) warming/greenhouse event rather than a cooling as has been previously interpreted. Moreover, sedimentary evidence, from the Sebree Trough and its southern margin in central Kentucky, shows that evidence for vigorous upwelling and significant nutrient-loading is lacking until the end of the GICE.

- Phosphatic limestones become more abundant for the first time in the base of the Grier Formation during the lowered sea-level and into the next transgression event. Thus, this pattern of increased upwelling (and transition back to lighter $\delta^{13}C$ values) would be expected if global cooling reinvigorated thermohaline circulation and upwelling reintroduced light carbon into the Sebree trough and onto the platform.
- Coincident with the GICE is the first pronounced, siliciclastic episode of the Chatfieldian (Trenton Group). Enhanced sedimentation, and nutrient loading, in the foreland basin, close to the orogenic center, would be expected and would account for greater fractionation and therefore heavier (more positive) isotopic values in this region.
- As subsequent, but smaller amplitude, carbon excursions seem to follow the same pattern relative to sea-level change, it is likely that these later excursions also record climate fluctuations (warming and cooling) that resulted in sluggish followed by more intense oceanographic circulation.
- Warming would favor increased partitioning/stratification of water masses within the Iapetus Ocean and marginal seas. With strong internal water-mass boundaries, it would be possible to isolate deeper-water environments (Sebree

trough and Taconic foredeep, for instance) from shallower ones (Lexington platform, etc.)

  - Hence it is suggested that sea-level oscillations (and carbon isotopic excursions) of the Chatfieldian (and perhaps earlier) may have been driven by glacio-eustatic processes.

- Strontium isotopes

  i. Strontium isotope ratios are primarily governed by changes in relative contribution of mafic-sourced strontium (lowered $^{87}Sr/^{86}Sr$ values) relative to felsic-sourced strontium (elevated $^{87}Sr/^{86}Sr$ values). These values are also influenced by hydrothermal delivery of Sr through the sea-floor.

     - Typically lowered (more negative) values are thought to reflect increased periods of seafloor spreading (increased hydrothermal leaching of Sr) or possibly strong weathering of large volumes of basalt.
     - Elevated (more positive) values likewise reflect weathering of continental crust (granites and/or average metamorphic rocks).
     - Although minimal (3x less than former), hydrothermal delivery of Sr through seafloor could either increase or decrease $^{87}Sr/^{86}Sr$ values. During most of the Ashbyan-Mohawkian this would tend to decrease $^{87}Sr/^{86}Sr$ values, although by late Chatfieldian this could elevate $^{87}Sr/^{86}Sr$ values.

  ii. Strontium isotopes $^{87}Sr/^{86}Sr$ have been shown to display a long-term decline during the Ordovician showing a significant change in cycling of strontium in the global oceans. This trend is documented, tentatively, to be interrupted by at least four prominent paired excursions (declining excursion followed by positive excursion) within the Ashbyan to earliest Edenian interval.

- Sr Excursion 1 brackets the transition between the Chazy and Black River groups with declining $^{87}Sr/^{86}Sr$ ratios during deepening and increasing $^{87}Sr/^{86}Sr$ ratios during shallowing and restriction.
- Sr Excursion 2 brackets the Turinian-Chatfieldian boundary. Declining $^{87}Sr/^{86}Sr$ ratios are observed in the interval containing the Millbrig K-bentonite although declining values initiate slightly earlier within rocks that show general shallowing. Rapidly, increasing $^{87}Sr/^{86}Sr$ ratios are demonstrated during the Chatfieldian (Rocklandian) during transgression and highstand. Equilibrium trend values are again achieved by the Kirkfieldian.
- Sr Excursion 3 occurs in the later Chatfieldian (*O. ruedemani* zone). It is less prominent than the first two events but shows declining $^{87}Sr/^{86}Sr$ ratios during deepening and increasing $^{87}Sr/^{86}Sr$ ratios during shallowing.
- Sr Excursion 4 occurs within the Edenian (latest *C. spiniferus* to *G. pygmaeus* zone) but is somewhat more prominent than excursion 3. Nonetheless, the excursion shows declining $^{87}Sr/^{86}Sr$ ratios during the last deepening of the Trenton Group and increasing $^{87}Sr/^{86}Sr$ ratios during highstand and subsequent shallowing of the Kope Formation and its equivalents including the Indian Castle Shales.

iii. Sr Excursions 1, 3, and 4 appear to be in phase with sea-level change

- Declining $^{87}Sr/^{86}Sr$ values during transgression thus likely reflect more intense periods of sea-floor spreading (SFS). It is expected that increased SFS might produce slightly more elevated sea-levels.
- Increasing $^{87}Sr/^{86}Sr$ values likely correlate with major increases in the delivery of siliciclastic sediments to the ocean as a function of sea-level drop and increased exposure area whereby at least some sediment could be transported by aeolian processes. Alternatively increasing $^{87}Sr/^{86}Sr$ values could reflect a strong tectonic pulse whereby continental crust and continental crust-derived sediments were uplifted and subjected to intense weathering.
- As suggested by carbon and strontium isotopes, deepening in the Late Ordovician may have been tied to warming events, which were in turn tied to increased rates in seafloor spreading, increased CO_2 output and therefore an increased greenhouse effect.

iv. The anomalous Sr Excursion 2, coincides with the interval containing abundant K-bentonites of the "Hagan K-bentonite swarm" which includes the extremely widespread and thick Deicke and Millbrig metabentonites.

- The strongly negative phase of the Sr excursion initiates in the interval containing the swarm and despite a phase of shallowing, Sr isotopic ratios suggest another major pulse in sea-floor spreading.
- Given the increased incidence of closely-spaced, and extremely voluminous volcanic eruptions at this time, it is possible that Sr Excursion 2 reflects a major pulse not only in SFS but also in subduction and partial melting, and may even

signal rapid melting after slab break-off during subduction reversal (as suggested by Karabinos et al., 1998).

    - Extreme shallowing, and rapidly oscillating sea-level, along with multiple extinctions both in macro and microbiota at this time, may reflect a pronounced volatility in climate. Given lowered sea-levels, despite intense volcanic degassing of greenhouse gases, it may have been possible that the eruptions produced a period of time with substantial cooling at the end of the Turinian. This was then followed by warming going into the early Chatfieldian.

- Neodymium isotopes

  i. Neodymium isotopes are proving extremely useful for provenance studies whereby the age of neodymium-bearing sediments might help to identify changes in the source and timing of sediments during foreland basin development (see Holmden et al., 1998). Neodymium isotopic ratios (εNd – relative to a standard) generally reflect the age of the source rock supplying the neodymium.

    - ^{144}Nd is a stable form, while ^{143}Nd is a decay product of samarium and increases in concentration relative to the stable form.
    - Therefore εNd becomes more negative with significantly older source rocks.
    - In marine environments, Nd is rapidly mixed and has a short residence time but is not typically homogenized in the global ocean. Instead, water masses, and especially shallow marginal and epicontinental seas, inherit a local Nd signature.
    - For much of the Ordovician, Laurentian rocks typically had εNd signatures characteristic of much older rocks relative to other cratons. However during

major transgressive events (i.e. the late Sauk sequence, immediately preceding the Knox Unconformity, and in the latest Ordovician Tippecanoe sequence preceding the Hirnantian), Laurentian εNd signatures become younger and more similar to global values as old, cratonic shield sediment source terranes are flooded and the supply of sediment decreases substantially.

- During major lowstands, including the Knox Unconformity, εNd signatures show a pronounced isolation from global signatures and significantly older source rocks dominate Laurentian signatures. Therefore εNd signatures reflect pronounced sea-level changes as sediment source terranes are flooded and subsequently exposed.

ii. At higher-resolution, during the Turinian to early Chatfieldian, εNd signatures show a pronounced high-order fluctuation in rocks of the Upper Mississippi Valley near the Transcontinental Arch. In this case, εNd values also vary with sea-level change.

- Typically more negative εNd values indicate significant exposure of old sedimentary terranes during regressive periods.
- Increased εNd values reflect submergence of source terranes and movement toward a younger sediment source terrane which in the Turinian to Chatfieldian was likely constrained to the Taconic Highlands.
- Some six excursions (I1-I6) have now been recognized in the Upper Mississippi Valley and co-vary with the major carbon isotopic excursions of the Chatfieldian. The oldest, I1, is coincident with the GICE and Sr excursion 2, and shows a pronounced "younging" of εNd values immediately after

deposition of the Elkport K-bentonite during a significant and pronounced sea-level highstand event.

- The “younging” has been attributed to the loss of Transcontinental Arch-derived sediments and an influx of Nd from more eastern sources (Grenville and Taconic Orogeny). Subsequent decrease in εNd values suggests sea-level lowering and an associated increase in exposure of local sediment source terranes.

iii. The coincidence of the pronounced strontium isotopic excursion with the I1 neodymium excursion collectively lend significant support to the premise that the first major pulse of siliciclastics in the early Chatfieldian was likely timed with a pronounced pulse in tectonism in the Vermontian tectophase.

- This tectonic pulse, timed with some of the first seismites of the Chatfieldian, formation of the Sebree Trough and other features, obviously involved significant faulting, thrusting, and uplift of a major terrane along the eastern margin of Laurentia. This terrane then began to shed significant quantities of sediments into the Taconic Foreland Basin and adjacent platforms.
- Nonetheless, the transportation of sediments some 2000 km inboard of the continental margin required a significant sediment transport mechanism, either transportation via internal water mass boundary layers, or possibly as wind-blown sediment.
- As suggested above, it is likely that warming may have produced pronounced stratification of foreland and epicontinental seas capable of long-distance sediment transport, i.e. along a pronounced pycnocline. Likewise, warming, a

sub-tropical paleogeographic latitude, and likely paleowind directions positioned eastern Laurentia so that it may also have received sediments transported by aeolian processes from the Taconic hinterland.

  - Obviously this event was the most pronounced and most significant of the Chatfieldian given the synchronicity of these major isotopic systems. Subsequent events were less so. Immediately following, and even during the early Chatfieldian events, the architecture of the GACB region was modified so significantly by block faulting that the unique oceanographic conditions that developed during the first events became less pronounced thereafter.

c) Collectively, the integration of these data suggest that sea-level change during the Late Ordovician (Mohawkian) may have been strongly tied to glacio-eustatic sea-level change with circa 1my periodicities. Moreover, major lithologic transitions, i.e. Chazy-Black River, Black River – Trenton, and Trenton-Utica/Kope represented pronounced pulses in tectonic events that altered the architecture, and sediment supply of the GACB. In addition, it is clear that the Turinian-Chatfieldian boundary interval, with significant changes in ocean chemistry, volatile sea-level, intensity of volcanic eruptions, all coincident with a number of extinctions represents an even more significant event that may be related to a reversal in subduction from east dipping to west dipping (see discussion Chapter 2).

CHAPTER 7: INTEGRATED SEQUENCE STRATIGRAPHIC FRAMEWORK OF ASHBYAN-MOHAWKIAN (UPPER ORDOVICIAN) STRATA OF EASTERN LAURENTIA

ABSTRACT

During the Late Ordovician (mid-late Mohawkian), the "Great American Carbonate Bank" (GACB) of eastern Laurentia became increasingly segmented due to the impact of the Taconic Orogeny. This same time interval records one of the great evolutionary radiation events of the Paleozoic, is thought to record a significant climatic change, and is followed by a significant mass extinction at the end of the Ordovician. The lack of a regionally-calibrated geochronology has hindered studies that might investigate the regional pattern of evolution with respect to climate and/or tectonic change on shorter time scales. The construction of a detailed geochronology, as documented herein, will enable a better understanding of how tectonic episodes impacted the GACB and its biotas.

This investigation focused on development of a sequence stratigraphic calibrated chronostratigraphic model for the entire Chazy-Black River-Trenton Group interval for key sites in eastern North America. Sites were selected based on rock exposure, availability of rock core, and historical precedent. Herein, 13 third-order depositional sequences (~ one million year duration) have been identified and correlated between New York/Ontario, central Pennsylvania, and Kentucky/Ohio. Sequence nomenclature expands on that of Holland and Patzkowsky (1996, 1998) and Brett and colleagues (2004). This work also establishes a tentative, detailed correlation with sequences described from the Late Ordovician of the Upper Mississippi Valley by Witzke and Bunker (1996).

Detailed comparisons indicate the importance of allocyclic controls on the development of sedimentary cycles. Nonetheless, important changes in sequence architecture are noted and

can be attributed to local to regional modification of the GACB by autocyclic phenomena beginning prior to or at least coincident with development of the Knox Unconformity. Protected, low-energy carbonate deposition on the GACB platforms during the Chazy-Black River interval (Sequences M1A-M5A) may have been the result of tectonic activation (uplift) of the GACB platform margin in the early Ashbyan (M1A sequence). Eustatic fluctuation and sedimentation through the mid to late Turinian stage ameliorated the impact of topographic features that impacted deposition during the Ashbyan.

Depositional facies become exceptionally similar and widespread beginning with the M2 sequence (M3-M4A are the most similar overall). These sequences contain a significant number of "time-restricted facies" and unique marker intervals. Faunal patterns are comparable between most regions and indicate periods of restriction and more open-marine circulation. These likely reflect allocyclic, eustatic fluctuations and associated climatic changes during a period of major expansion in the "Black River Lagoon." This may have been tied to a period of increased sea-floor spreading rates which resulted in eruption of the large-scale Deicke and Millbrig K-bentonites in the M4A-M4B sequences respectively.

Substantive changes were initiated during the M5A sequence through the introduction of siliciclastics across the entire GACB with significant topographic contrasts that initiated after deposition of the M5B sequence. Depositional sequences are identifiable in late Chatfieldian (Shermanian) strata and are most comparable between New York and Kentucky as noted previously. Nonetheless these sequences, and especially the M5C, M6A, M6B, and M6C, show substantially more variability in local areas than do older sequences. Most often these are manifest by synchronous, abrupt lateral facies change, and evidence for synsedimentary deformation across widespread areas (including in all three study areas). It remains to be seen if similar synsedimentary deformation/faulting events can be recognized in the Upper Mississippi Valley region. Nonetheless, given their recognition in the Nashville Dome, the Jessamine Dome, the Pennsylvania Embayment, and the New York Platform, it is highly likely that these imply tectonism as a major control, not only on the architecture of localized regions of the Late Ordovician carbonate platforms of eastern Laurentia, but also on their final demise during the earliest Edenian (C1 sequence).

INTRODUCTION:

Strata of the Upper Ordovician of eastern Laurentia have become an important testing ground for application and refinement of sequence stratigraphic concepts as applied to portions of the Taconic foreland basins and their adjacent margins (Holland & Patzkowsky, 1996; 1997; 1998; Pope & Read, 1997a, 1997b; Hohman, 1998; Joy et al., 2000; Cornell, 2001; Brett et al., 2002; Brett et al., 2004; McLaughlin, et al., 2004, McLaughlin 2007). This work has built on previous outcrop-level syntheses in order to establish a framework of lithostratigraphic, biostratigraphic, and event stratigraphic horizons within strata from distinct regions of eastern Laurentia. Research has also focused on recognition and correlation of unique horizons, including K-bentonites and various isotope excursions (Haynes, 1994; Huff et al., 1992; Kolata et al., 1996; Patzkowsky et al., 1997; Fanton et al., 2002; Saltzman et al., 2003; Ludvigson et al., 2004; Mitchell et al., 2004; Barta, 2005; Panchuk et al., 2005; Young et al., 2005; Fanton & Holmden, 2007). The nature and diversity of these studies have provided a plethora of independent, long-distance correlation tools that have been linked elsewhere in this dissertation for linking stratigraphic frameworks from different regions of the Laurentian Epicontinental Sea and the adjacent foreland basins. Collectively these analyses are making it possible to build a refined, basin-wide, chronostratigraphic framework for the Great American Carbonate Bank and provide insights for evaluating the nature of its demise during one of the most important periods of biotic and environmental change in the Phanerozoic.

Nonetheless, to date, only preliminary efforts have been made to link portions of the Great American Carbonate Bank (GACB) using sequence stratigraphy in an effort to investigate the relative impact of the Taconic Orogeny on the GACB (see Holland & Patzkowsky for the Cincinnati Arch, 1996; and Brett et al., 2004 for the Cincinnati Arch and New York - Ontario). This is primarily due to the interplay between tectonics and eustasy during the activation of the passive margin and the perceived local to regional tectonic overprinting of facies change and expected diachroneity that makes sequence stratigraphic approaches challenging (i.e. see Joy et al., 2000). It is becoming clear that previous correlations and assumptions are still in need of additional refinement as are the details of specific K-bentonite correlations in particular.

However, significant progress has been made toward development of a sequence framework for this interval. Sequence stratigraphic principles have been re-evaluated and tested in the context of mixed carbonate-siliciclastic epicontinental seas (Brett et al., 2004, McLaughlin et al., 2004, McLaughlin & Brett, 2007, McLaughlin, 2008). These authors have used the Lexington Platform/Jessamine Dome region as a laboratory for identification and correlation of important stratigraphic surfaces, systems tracts, and associated patterns of biostratigraphic and lithologic change during the transition from carbonate-dominated to siliciclastic-dominated deposition. For the first time, robust, high-resolution sequence stratigraphic models of the Lexington Limestone have been established and correlated for distances of up to 300 km perpendicular to depositional strike.

McLaughlin and colleagues (2004) used closely-spaced outcrop sections and numerous subsurface cores to construct a high-order framework of correlations from relatively shallow-water environments of the southern Jessamine Dome northward into relatively deep-water environments of the Sebree Trough. This study, and subsequent studies (i.e. McLaughlin and Brett, 2007) investigated: 1) changes in facies distribution patterns, 2) physical event horizons (including K-bentonites, soft-sediment deformation intervals, etc.), 3) occurrence and depositional characteristics of hardgrounds and other diastems, and 4) bioevents and unique taphonomic events using faunal-gradient analysis . This work has been critical in developing a methodology that builds off previous investigations and integrates seemingly disparate sedimentologic, stratigraphic, and paleontologic approaches for the purpose of basin analysis.

The current study is thus focused on extending the sequence stratigraphic methodology and assessments of the Jessamine Dome region into New York and Ontario as well as into central Pennsylvania's Ridge and Valley outcrop region. Herein the focus is on strengthening the chronostratigraphic framework of previous authors, including the pioneering work of Holland & Patzkowsky (1996, 1998) and building upon the initial analysis of Brett and colleagues (2004). The ultimate goal is to provide a detailed chronostratigraphic model for the entire GACB platform, platform margin, and the developing foreland basin in order to investigate the unique interplay between tectonics and eustasy during the different tectophases of the Taconic Orogeny. In order to understand the complexity of the Taconic Orogeny and its associated tectophases, especially the Blountian and Vermontian tectophases, this work, presents an expanded look at the sequence stratigraphy of limestones of the Chazy, Black River, and Trenton groups and their transition during two-phases of siliciclastic-dominated deposition during the Ashbyan to Shermanian Series (~460.5 mya to 450.5 mya).

SEQUENCE STRATIGRAPHIC HIERARCHY & RESEARCH PROBLEM

Sequence stratigraphy is a methodology used to establish or predict the spatial patterns of deposition of a constrained sedimentary succession deposited during episodes of sea-level rise and fall. It is typically defined as a stratigraphic method that uses unconformities (sequence boundaries) and their correlative conformities to package sedimentary successions into spatially and temporally constrained units: sequences (Vail et al., 1991; Vincent et al., 1998; Emery & Myers, 1996). Any unconformity bounded "sequence" can then be divided internally into smaller-scale genetically related units (i.e. systems tracts). These are deposited during individual phases of sea-level change (i.e., transgression, highstand, regression, and lowstand). This method is a powerful tool in modern stratigraphic studies because it integrates many aspects of stratigraphy including seismic stratigraphy, lithostratigraphy, cyclostratigraphy, event-stratigraphy, and biostratigraphy into a single predictive framework. The development of a sequence stratigraphic framework for any given depositional basin provides the stratigrapher not only with a temporal framework for studying depositional change, but it also provides a spatial framework.

An important aspect of sequence stratigraphy is its use in basin analysis through the establishment of very low-resolution (megasequence scale) to very high-resolution spatio-temporal patterns (parasequence scale). There are typically a number of different “orders” of sequences observed within the Ordovician. The magnitude of such sequences, as discussed by a number of authors including Van Wagoner (1988), and Vail and colleagues (1991), generally is categorized into a variety of scales of temporally and spatially constrained depositional sequences. Those deposited during long time scales (i.e. 100 million years) and with large sedimentary thicknesses (1000's of meters) are generally referred to as 1st-order tectono-sequences. Within these tectono-sequences are slightly thinner "sequences" of slightly shorter durations and are similar in scale to the "megasequences" of Sloss (1963) – these are the typical 2nd order super sequences. These are in turn divisible into a number of 3rd, 4th, 5th-order, etc. sequences that are developed over shorter time-scales, and range in thickness from 100's of meters to a few meters or less in the case of the higher-orders (**Table 1**).

Order	**Approximate Duration (millions of years)**	**Approximate Thickness**	**Relative sea-level amplitude (meters)**	**Sequence Nomenclature**
First (1rst-order)	>100	1000's of meters		Tectono-Sequence "Wilson Cycle"
Second (2nd-order)	10-100	100's of meters	50-100	"Supersequence" Sloss's Megasequences
Third (3rd-order)	1-10	10's of meters	50-100	Depositional Sequence Composite Sequences
Fourth (4th-order)	.1-1	meters	1-150	High Frequency Sequence Parasequence & Cycle Set
Fifth (5th-order)	.01-.1	10's of centimeters	1-150	Parasequence High-Frequency Cycle
Sixth (6th-order)	.001-.01	centimeters		Hemicycles?

Table 1: Orders of stratigraphic sequences as applied by Goldhammer (2003) (from Encyclopedia of Sediments and Sedimentary Rocks, Middleton et al., eds. 2003).

It is clear from the sedimentary record on the North American craton, that Sloss's (1963) Tippecanoe Megasequence represents one of the highest sea-levels ever recorded in the Phanerozoic. Ranging in duration from the Middle Ordovician to the end of the Early Devonian (some 80 to 90 million years), the development of the Tippecanoe Megasequence represents one of the longest periods of sea-level highs thus far recorded. However, the Tippecanoe megasequence was not deposited by a single large-scale, long-term sea-level rise and fall event. Instead, it was punctuated by the development of a series of additional unconformities effective on a variety of shorter and narrow temporal and spatial scales. The subject of this study focuses on the earlier portion of the Tippecanoe Megasequence (Creek Holostrome; after Wheeler, 1963), which is constrained by the Knox Unconformity below and by the Cherokee Unconformity at the Ordovician/Silurian Systems boundary. Superimposed on this succession are the higher frequency sequences previously recognized by Pope and Read (1997, 1998) and by Holland and Patzkowsky (1996, 1997, 1998; see below). These record much shorter duration sea-level events, that when considered collectively produce the patterns associated with the larger-scale megasequences. Nonetheless, within this succession of sequences, are a number of local to regional changes in sedimentary regime. These changes reflect a complex hierarchy of biotic, oceanographic, and climatic events that are certainly inter-related.

However, the largest problem faced in the development of a comprehensive understanding of Taconian tectonism and its impact on eastern Laurentia during the Upper Ordovician, has been the lack of high-resolution, inter-regional correlations. In order to affect a more refined geochronology, previous studies have been integrated herein into a high-resolution framework of relatively narrow time slices using bio-, litho-, and event stratigraphy as outlined in detail in other parts of this dissertation. Building on the initial work presented by McLaughlin and Brett (2004), and Brett and colleagues (2004), these time slices and depositional sequences have been used to establish an extended sequence stratigraphic framework for the entire Ashbyan to Mohawkian interval of the Cincinnati Arch, the type Mohawkian interval in New York-Ontario and the carbonate-clastic succession in central Pennsylvania. This compendium builds upon the observations and analyses of numerous local to sub-regional studies and extensive outcrop and core investigation, to develop a well-constrained, framework of marker horizons and time equivalent sedimentary packages across 1000's of square kilometers of the eastern United States and Canada.

REVIEW OF MOHAWKIAN SEQUENCES

Several studies have investigated and delimited a number of depositional sequences within the Upper Ordovician strata of the Cincinnati Arch region. Pope and Read (1997a, 1997b; 1998) subdivided Ordovician rocks of the Cincinnati Arch to Virginia region into three second-order "supersequences" (designated supersequences 1, 2, and 3) of roughly ten to thirty million year durations. Their second and third supersequences (also referred to as the "Blountian" and "Taconic" supersequences respectively) include the interval from the base of the Ashbyan (Knox Unconformity) to the end-Ordovician unconformity – thus coinciding with the Creek Holostrome of the Tippecanoe megasequence of Sloss (1963; **figure 1**). The Blountian-Taconic

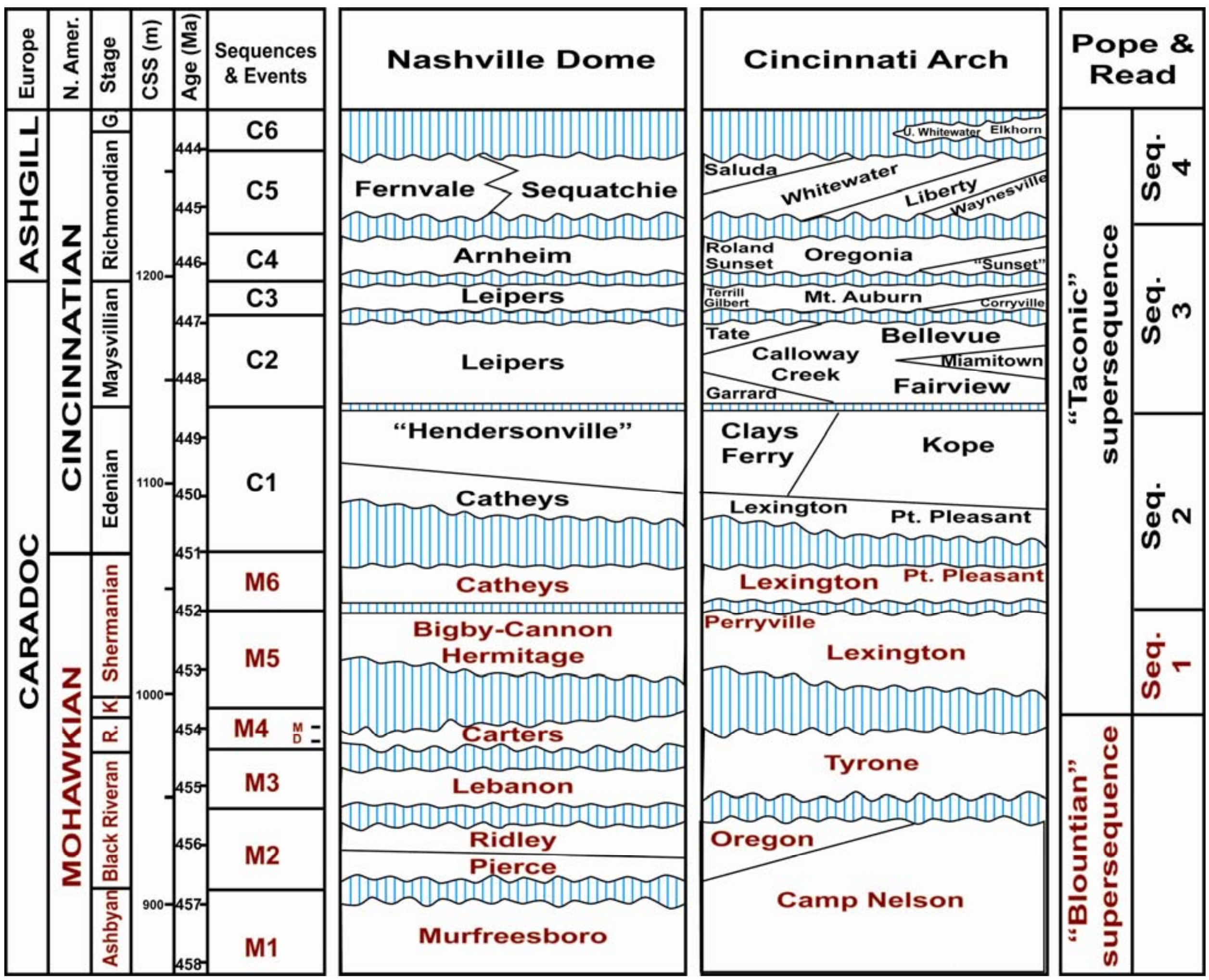


Figure 1: Sequence architecture for the Upper Ordovician as established by Holland & Patzkowsky (1997, 1998, 2008) and by Pope & Read (1997, 1998). Third-order depositional sequences outlined by Holland and Patzkowsky(i.e. M1, M2,

etc.) do not generally coincide with those established by Pope and Read (Seq. 1, Seq.2, etc.), except see Sequence 1 and the M5 sequence that do coincide. Sequences investigated in this study are shown in dark red.

supersequence boundary was placed at the base of the Lexington Limestone and just below the base of the Martinsburg Formation (& uppermost Eggleston Formation) of Virginia (Pope et al., 1997). Pope and Read (1997, 1998) further subdivided their large-scale sequences into several third-order depositional sequences and each of these was in turn, divided into meter-scale cycles. As shown on figure 1, these sequences are labeled Seq. 1 to Seq. 4 and are considered by Pope and Read (1998) to represent durations of one to ten million years each.

Likewise, Holland & Patzkowsky (1998) identified eleven large-scale (third-order) depositional sequences in the Middle and Upper Ordovician of the Nashville Dome (now all included in the Upper Ordovician). These sequences were developed from the Ashbyan (earliest Mohawkian) upward to near the top of the Richmondian (latest Cincinnatian). These were developed through an interval of approximately fourteen million years (458 mya to ~444 mya). In their study, the Mohawkian succession was divided into six large-scale depositional sequences (M1 through M6), each with a duration ranging between one to three million years. The ensuing Cincinnatian was also defined to have six depositional sequences (C1-C6), although C6 was not present in the Nashville Dome region. Thus, in comparison to Pope and Read (1998), Holland and Patzkowsky identified eight as opposed to four third-order depositional sequences in the "Taconic" supersequence with differing stratigraphic boundaries. This indicates significant differences in how depositional sequences were assigned. As suggested by McLaughlin and Brett (2007), it appears that the regularly repeating pattern of facies change reflective of depositional sequence development in this region was not consistently recognized by these former authors and was not well-tied to the stratigraphic pattern present in these rocks.

Subsequently, McLaughlin and Brett (2004), and Brett and colleagues (2004) subdivided the succession of Mohawkian sequences, first recognized by Holland and Patzkowsky, into regularly repeating, stratigraphic intervals of slightly shorter duration. The M5 and M6 depositional sequence were thus sub-divided into three depositional sequences each (M5A, M5B, M5C, and M6A, M6B, M6C) that reflected the original architecture proposed by Holland and Patzkowsky with only slight modification of the M6 to C1 sequence boundaries. McLaughlin and Brett showed these sequences were well-developed and correlated across the Cincinnati Arch region. The lower portions of these depositional sequences showed prominent, relatively clean, coarse-grained limestone facies with deepening-upward patterns (often capped by distinct hardgrounds and condensed intervals). Each of these was in turn followed by gradual shallowing upward intervals that were characteristically siliciclastic-rich and often contained calcareous and/or organic-rich shales interbedded with deep-water carbonates, forming allodapic rhythmite facies.

Hence, the M4, M5A, M5B, M5C, M6A, M6B, M6C, and C1 sequences (upper High Bridge Group, Lexington Limestone, and basal Cincinnati Group) of the Cincinnati Arch region were investigated and correlated in great detail in both outcrop and cores using litho-, bio-, and event stratigraphy (McLaughlin et al., 2004; McLaughlin and Brett, 2004). The details of these sequences were then correlated into the New York – Ontario type region by Brett and colleagues (2004; **figure 2**). Detailed comparisons between New York and Kentucky have revealed similar patterns in facies development, sequence architectures, and timing of basin subsidence. Both large-scale sequences and smaller-scale cycles initially are: 1) widely correlatable (Chazy – Black River Sequences M1-M5A), 2) show synchronous water depth histories, and 3) gradual lateral changes in shallow facies suggesting broad facies belts and extensive flat-topped platform architecture. Sequences and cycles appeared to respond to high-order sea-level fluctuations

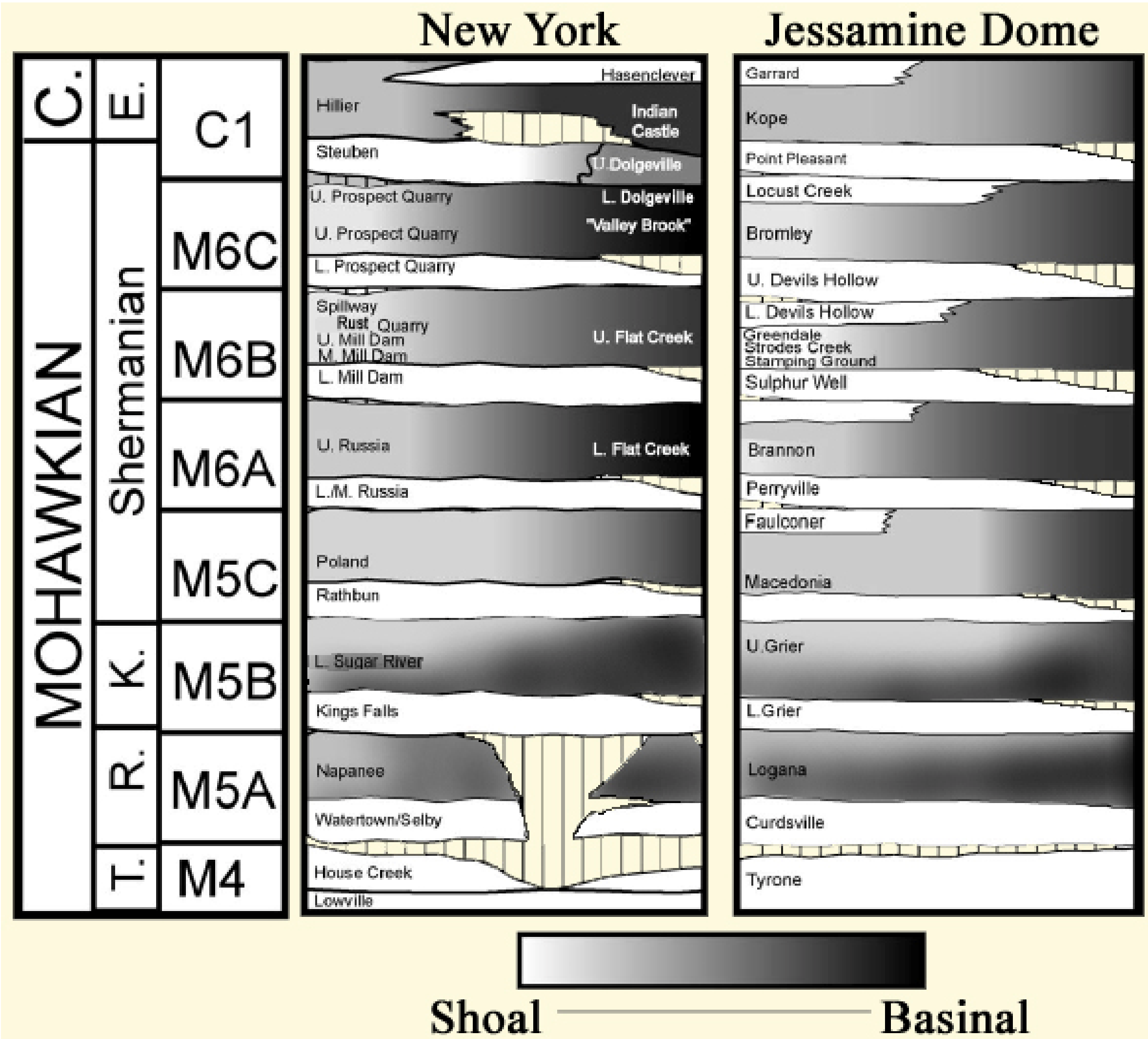


Figure 2: Depositional sequences and stratigraphic comparison of the Mohawkian to basal Cincinnatian interval of the Jessamine Dome (Cincinnati Arch) and the New York region (after Brett et al., 2004).

suggesting that eustasy played a major role in their deposition. Conversely, sequences of the mid-to-late Trenton (M5B to C1) record abrupt lateral facies changes from shallow, peritidal carbonates into deeper water nodular carbonates and shales. This significant change is indicative of extensive lateral topographic modification of the platform after the onset of tectonism.

SEQUENCE MOTIFS

Sequences, M1 through M6 of Holland and Patzkowsky (1997, 1998) are the focus of this study. In their study, depositional sequences were generally recognized to contain transgressive systems tracts and highstand systems tracts, and all but one lacked evidence for lowstand systems tracts (Holland & Patzkowsky, 1998). These authors recognized that many of the sequences display evidence of subaerial exposure, including possible paleokarst development, and erosional truncation at the level of sequence boundaries. Other prominent features were also identified and included hardgrounds, maximum flooding zones, and extensive evidence for condensed intervals at or near the transgressive surface. However, as these authors duly note, the expression of depositional sequences is marked by a significant and pronounced change during the onset of the Vermontian tectophase of the Orogeny. Ultimately, it appears that although the Nashville Dome region was proximate to the Sevier Foreland Basin, foreland basin development of this region during the Blountian Tectophase was rather limited. Thus Holland & Patzkowsky recognize the Mohawkian sequences and their parasequences to reflect two different depositional motifs, the first set characteristic of the “pre-Taconic platform” (or pre-Vermontian) and the “post-Taconic platform” reflective of the nomenclature used by Pope and Read (1997a,b).

The first group of sequences (M1 – M4), were generally characterized by a wide array of off-shore through supra-tidal facies. These sequences are relatively fine-grained, micrite-dominated limestones deposited in a range of protected, low-energy depositional environments. Most are dominated by massive-bedded, peloidal micrites with occasional skeletal wackestones and packstone facies. The latter of which, typically show extensive bioturbation and contain abundant *Chondrites* and *Phycoides* traces. These are variously interbedded in shallowing-upward motifs that culminate in desiccation-cracked micrites exhibiting tabular laminae and widespread fenestral fabrics. In some sequences, these are capped by highly dolomitized intervals produced either as a result of primary dolomitization or as early secondary dolomitization. Based on correlation of these facies laterally, it was inferred that facies belts were broad and extensive reflecting a flat-topped platform environment over most of the Nashville Dome region, and, by extension, the Jessamine Dome to the north (Holland & Patzkowsky, 1996). The lack of appreciable coarse-grained, grainstone facies in these sequences indicated that high-energy, barrier shoal facies were not present or if they were present must have existed outside of the region. Alternatively, based on its broad, shallow architecture, the carbonate platform may have prevented development of highly energetic environments.

The second set of sequences, the “post-Taconic platform” sequences, was dominated by deep subtidal to low-energy, shallow sub-tidal, and high-energy, sand-shoal environments. These sequences rarely developed the peritidal facies that had been so dominant and widespread in the earlier sequences (Holland & Patzkowsky, 1997a,b; 1998). Instead the M5-C5 sequences are dominated by generally deeper-water lithologies influenced by significant siliciclastic sedimentation. In addition, these sequences contain deeper-water faunas, and a wider array of lithologic types that grade laterally into deeper-water shaly facies over relatively short distances. Collectively these observations suggested to Holland and Patzkowsky (1996, 1998) that the Nashville Dome and by extension the Cincinnati Arch, experienced a major reorganization of platform architecture as a direct, distal orogenic effect of the Taconic Orogeny. The effects of this change were first recorded in their M5 sequence. Moreover, these authors supported the idea that the Nashville Dome may have become a significant feature as a peripheral uplift or bulge adjacent to the newly developed Taconic foreland basin at this time.

Holland & Patzkowsky (1996; 1997, 1998) noted the following as evidence of the shift in depositional regime: 1) the significant increase in siliciclastic influx onto the GACB (their Nashville Dome) during sequence M5, 2) associated changes in faunal assemblages, 3) a change in carbonate depositional regime from tropical to temperate-style deposition, 4) transition from flat platform-style environments to ramp-style environments due to tectonic modification of the GACB, as well as 5) a significant increase in the load of nutrients including phosphate within the Taconic basin (**figure 3**).

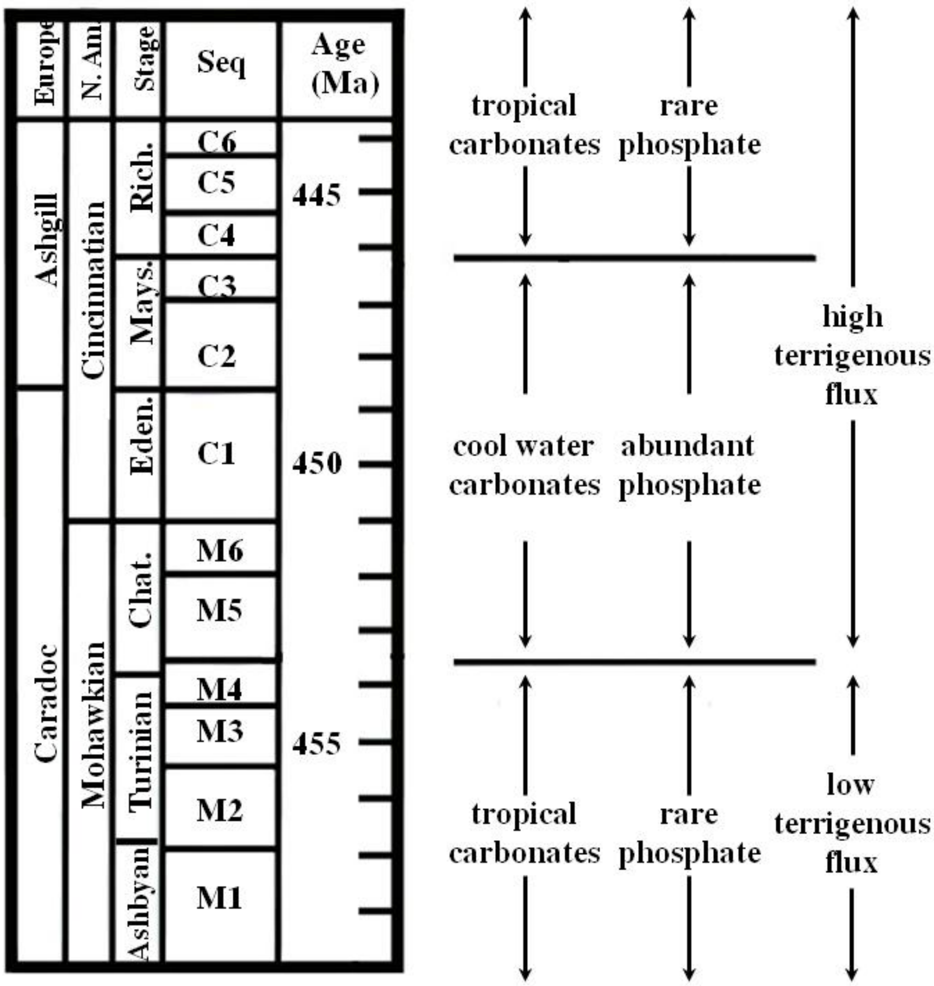


Figure 3: Depositional sequences for the upper Ordovician of the Nashville Dome of eastern North America (modified after Holland & Patzkowsky, 1996; 1998). Also shown are major sedimentologic changes observed in the Nashville Dome Region during the Taconic Orogeny.

Herein, thirteen, third-order depositional sequences for the Ashbyan to Mohawkian interval have now been constructed between two platform-to-ramp-to-basin transects in eastern Laurentia and compared to a stratigraphic succession deposited on the cratonic margin (**figure 4**). These transects are located in widely separated regions: a NW-SE cross-section of the carbonate bank to Taconic foredeep in New York State, and a S-N cross-section from the

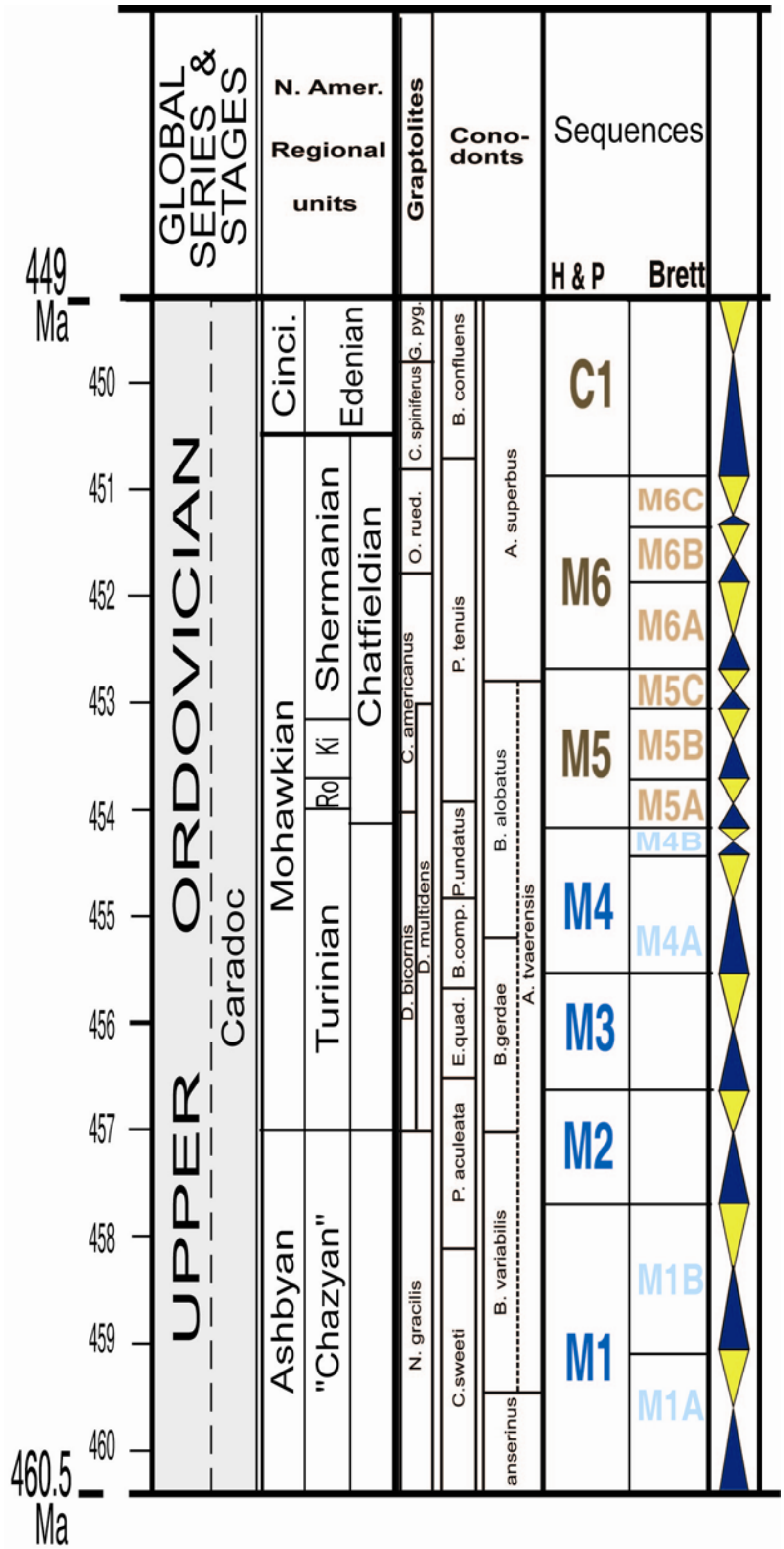


Figure 4: Sequence Stratigraphic and Biostratigraphic Framework for the Upper Ordovician of eastern Laurentia. Depositional sequences M1-C1 are after Holland & Patzkowsky (1996, 1998), while sub-sequences M5A-M6C are after McLaughlin et al., 2004, and Brett et al., 2004. Additional subsequences are now recognized within the M1-M4 interval and are documented here.

Lexington Platform into the Sebree intracratonic trough in Kentucky and Ohio. These sequences are compared in detail to the stratigraphic succession recorded in the Ridge and Valley of central Pennsylvania. Moreover, the succession of sequences is for the first time related to "transgressive-regressive cycles" (or T-R cycles) established in the Upper Mississippi Valley (Witzke & Bunker, 1996) in order to establish linkages with the type areas of several key marker intervals as discussed in previous chapters. As shown in **figure 4**, the sequence hierarchy established previously is modified to reflect additional depositional sequences newly recognized in the M1 and M4 sequences of Holland and Patzkowsky (1997, 1998).

SEQUENCE COMPONENTS

General sequence architecture

As mentioned, several orders of cyclicity are recognizable in the Upper Ordovician carbonate and mixed carbonate-siliciclastic successions of eastern North America. The smallest correlatable units are meter-scale cycles. These cycles, interpreted as parasequences (sensu Van Wagoner et al., 1988), generally show well-defined stacking patterns that overall show either deepening-upward (retrogradational), or shallowing-upward (progradational) trends. Cornell (2001) showed that numerous meter-scale cycles were easily correlated over wide areas within the Black River succession of New York and Ontario using among other things: overall cycle thickness, unique lithologies, and event beds (**figure 5**). Brett and Baird (2002) have similarly described and correlated a number of meter-scale cycles within the Trenton Limestone. In subtidal shelf facies these cycles commence with thin-bedded calcisiltites/lutites and shales and pass upward into bioturbated nodular to wavy-bedded wacke- and packstones and finally into amalgamated pack- and grainstones (Brett and Baird, 2002).

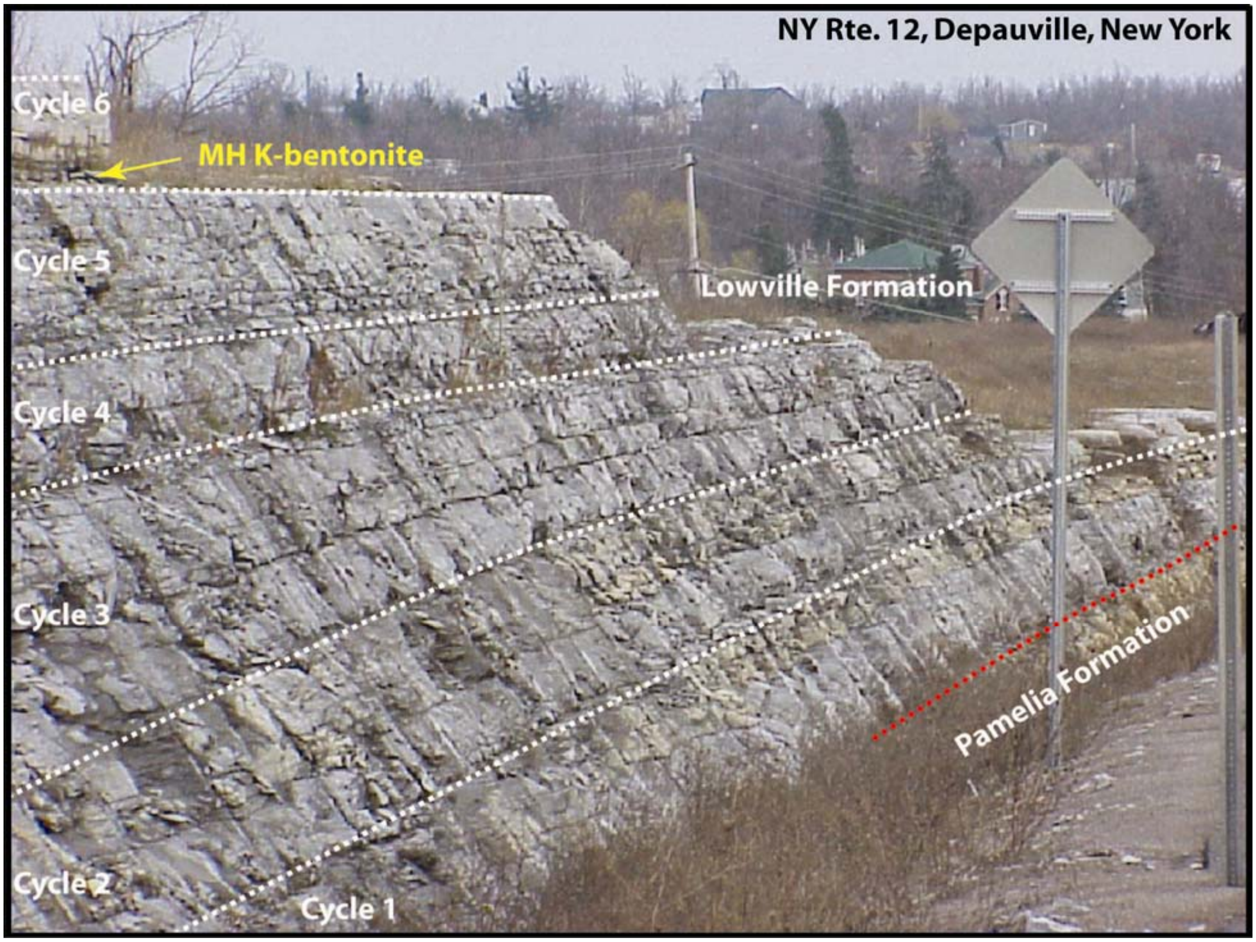


Figure 5: Outcrop photograph of the lower member of the Lowville Formation at Depauville, New York showing the position of six shallowing-upward cycles. Each cycle combined shows an overall transition from supratidal-capped cycles to shallowest sub-tidal to intertidal capped-cycles at the top of the succession suggesting an overall deepening pattern associated with a longer-term period of sea-level rise.

In addition to small-scale cycles or parasequences, thicker discontinuity-bounded depositional sequences of at least two orders of magnitude have also been recognized (Brett et al., 2004; McLaughlin, 2007). These sequences are typically five to fifteen meters-thick, and are thought to record depositional cycles of a few hundred thousand years duration (similar to fourth-order sequences of Vail et al. (1991). Decameter-scale cycles are in turn bundled in slightly thicker successions on the order of several tens of meters with temporal resolutions of up to about a million years. Typically these are considered third-order sequences (Vail et al., 1991). Third-order depositional sequences recognized by Brett and colleagues (2004), represent subdivisions of the composite third-order sequences (M5 and M6) recognized by Holland and

Patzkowsky (1996, 1998). Moreover, these are similar to the "parasequence sets" recognized by Pope and Read (1997a, b), but are interpreted differently (see Brett et al. 2004; McLaughlin et al., 2004).

Sequence Boundaries

One of the most important components of a depositional sequence are the sequence bounding surfaces that represent the change between shallowing-upward cycle stacking patterns and deepening-upward cycle stacking patterns or otherwise the lowest relative sea-level position. Thus, these bounding surfaces have also been referred to as maximum regressive surfaces (Catuneanu, 2006). In the Upper Ordovician, large, composite, and smaller scale sequence boundaries are recognized in most outcrops and cores as a relatively sharp contact that demonstrates evidence for erosion and/or a major change in facies. On occasion some of these are recognized as karstified surfaces where significantly shallower facies are deposited directly over deeper water facies. This facies dislocation is interpreted as a basinward shift in facies, across a sequence boundary (Emery and Myers, 1996). As indicated by McLaughlin and colleagues (2004), the sequence boundary is a subaerial unconformity in proximal areas and grades into a sharp, typically planar, submarine erosion surface in more distal areas above normal wave base. In the latter case, wave agitation is argued to bevel the seafloor to produce the flattened surface. In even more distal areas below normal wave base, the sequence boundary becomes a correlative conformity and maybe very difficult to place (Van Wagoner et al., 1990; Emery & Myers, 1996).

In some cases, sequence boundary contacts may be paired with a parallel but higher transgressive surface. More commonly they are combined to form erosion/transgression (E/T) surfaces. In such cases, the sequence boundary and the overlying transgressive ravinement surfaces are often merged to form one composite surface. Transgressive ravinement surfaces are developed during early transgression after the sequence boundary is formed, but as shoreline position begins to reverse from a progradational to retrogradational pattern. The result is reworking, winnowing, and/or redistribution of sediments deposited on the sequence boundary during the ensuing sea-level rise. Typically, these E/T surfaces are laterally extensive and regionally show some evidence for diachroneity. In most cases, however, on regional scales these surfaces can be shown to truncate underlying beds and can be recognized below intraformational conglomerates.

Lowstand deposits (LSTs)

As in the Nashville Dome (Holland & Patzkowsky, 1997), Chazy, Black River, and Trenton Group-facies on the platform of the Cincinnati Arch and New York-Ontario shelf show little evidence for lowstand deposits (LST). The LST, the stratigraphically oldest deposits of a depositional sequence, are deposited during the time when sea-level is moving from its lowest point through the ensuing earliest transgressive phase (Emery & Myers, 1996). LSTs are not typically found on carbonate platforms, but may be developed in deeper off-platform settings either as massive breccia fans resulting from destabilization of shelf margin facies during late stages of sea-level fall, or more typically as bundles of allodapic carbonates intercalated with dark shales. Such facies are observed in some portions of the GACB proximal to the foredeep basins associated with basin-forming tectophases events.

For example, significant numbers of carbonate shelf-derived intraformational breccias associated with deep-water facies are deposited first in mid Turinian sequences in the vicinity of the Sevier Basin. In this case, lowstand deposits are represented by the lower and upper Fincastle Conglomerates of Virginia which appear to have developed during at least two different sequences. More widespread lowstand intraformational breccias are found in early to mid Chatfieldian sequences along the margin of the Martinsburg Basin. In these cases, deposits including the “Paxton Creek” or “Hershey Conglomerate” of the Great Valley of east central Pennsylvania, the “Rysedorph Hill Conglomerate” of eastern New York, and others appear to have been deposited under such circumstances. It appears that these latter breccia deposits were subsequently incorporated into the thrust complexes associated with later phases of convergence and were additionally altered and fragmented.

The second facies that is diagnostic of LST deposition are the allodapic carbonate – shale rhythmite facies. These are typically much more difficult to separate from other sequence components in foreland basin settings because they are all typically deep-water deposits, with only minor variations in the carbonate to shale ratio. Nonetheless, there are a number of possible lowstand deposits recognized in foreland basin settings. These include, for example, portions of the Dolgeville Formation of the Mohawk Valley of New York (see Baird and Brett, 2002), as well as potentially portions of the Salona and Coburn Formations of central Pennsylvania.

Transgressive Systems Tract (TST)

Transgressive systems tracts (TSTs) form during interval when sea-level is rising rapidly to its maximum level. Typically in Mohawkian sequences two motifs for TSTs are developed and depend on platform architecture and relative sea-level position. In peritidal environments typical for instance of Black River to lower Trenton Group sequences, TSTs are recorded by quartzose to glauconitic lag beds and the transition from dolomitic limestones (occasionally evaporitic) and fenestral micrite facies to fossiliferous and bioturbated wackestones. Before the M5A sequence, these facies are very extensive and widespread reflecting broad-facies belts and relatively planar platform architectures in all of the study areas discussed here.

Beginning with the M5A sequence, peritidal facies can be traced down depositional dip into calcarenite and interbedded grainstone-rudstone facies (i.e. Watertown Limestone of New York, and or Salvisa-Perryville succession of Kentucky) across relatively short distances. This indicates a significant shift in platform architecture of the GACB to more steeply dipping ramps and narrower facies belts. Even, though siliciclastics become significant components of depositional environments in the M5A sequence, shallow, subtidal intervals within the Trenton are often composed of clean, widespread pelmatozoan-brachiopod pack to grainstones and are interpreted as TSTs. During deposition of TSTs, siliciclastic sediments are sequestered in nearshore estuarine regions and are not transported offshore or onto platforms devoid of siliciclastic source areas. Many of these skeletal limestones have been traditionally interpreted as the caps of shallowing upward parasequences (Pope & Read, 1998). Nonetheless, these pack- and grainstone facies show a taphonomic signature, especially in the middle to late TST, where the preservation of fossils is typically good. In such cases, numerous fossils including echinoderms, trilobites, and delicate brachiopods can be articulated and preserved intact and/or are typically very poorly sorted. In contrast, shallowing-upward intervals typically display highly abraded, well-winnowed and sorted cross-bedded calcarenites that are more typical of shallowing or Regressive Systems Tracts (RSTs).

While TSTs may appear massive and homogeneous, when examined in detail, TSTs are composed of multiple small-scale cycles that show shallowing-upward patterns and are typically demarcated individually by thin argillaceous to stylolitic partings. However, overall these stacked cycles show increased condensation and deepening-upward signatures and changes in faunal assemblages from more robust and massive forms to more delicate and diminutive taxa. Thus, these small, meter-scale parasequences exhibit subtle to more pronounced retrogradational stacking patterns both in terms of lithology and paleontology. These cycles are remarkably consistent in thickness laterally in shallow portions of the platform, and overlie sequence boundary surfaces. Predominantly within the Trenton, TST cycles show development of mineralized hardgrounds which may contain reworked litho and bio-clasts, evidence for mineralization including development of phosphatic, pyritic, glauconitic and/or chamositic grains, and/or exceptionally preserved hardground communities. These mineralized surfaces are especially more pronounced near the top of the TST where rates of flooding are most rapid and the effect of sediment starvation is most pronounced.

McLaughlin and colleagues (2004) recognized that in more basinal settings, typical of the "Taconic" sequences, TST facies are still recognized although they are typically much thinner than in shallower portions of the ramp and can be traced up ramp using K-bentonites. These deep-water TSTs are characteristically represented by brachiopod shell beds composed almost entirely of small dalmanellid brachiopods and/or by the inarticulate brachiopod *Leptobolus*. Ostracods and conodont lags are also typical in the most basinal settings along with the inarticulate brachiopods. These thin shell beds are often pronounced in core sections and show thin pyrite and phosphate-coated skeletal hash stringers especially close to the top of the TST.

<u>Maximum Flooding Surfaces (MFS)</u>

A second important surface within a depositional sequence is the Maximum Flooding Surface (MFS), also referred to as maximum starvation surfaces (MSS) by McLaughlin et al.,

2007. These surfaces form when sea-level is rising at its maximum rate and/or has reached maximum relative sea-level. In the Late Ordovician, maximum flooding surfaces in some cases are developed within a thin-interval and hence may be recognized as a zone of maximum flooding rather than a single surface. Typically, the true MFS or maximum flooding zone is underlain by a surface of maximum sediment starvation (SMS, sensu Baum and Vail, 1988). Above the SMS, the maximum flooding zone is typically a thin interval of calcareous shale and fine-grained, argillaceous limestones commonly with hardgrounds and corrosion surfaces. These thin intervals are considered to mark a condensed interval, with at least some evidence for condensation and mineralization denoted by thin concentrations of highly-fragmented fossil debris, phosphatic or pyritic debris, and organic matter enrichment. As such, these typically show very high gamma ray signatures on wireline logs, and are used to infer maximum flooding zones and the approximate position of the deepest water facies within a depositional sequence.

In Trenton sequences, these MFSs or maximum flooding zones stand out as an abrupt shift to condensed, fine-grained, deeper water facies overlying coarser-grained skeletal limestones of the underlying TST. In these sequences, the most pronounced change is the abrupt appearance of siliciclastic-rich facies at the base of the overlying highstand. In the pre-Trenton sequences, MFSs are typically less pronounced, but generally show a pause and/or reversal in the retrogradational stacking pattern.

Small, meter-scale cycles develop either aggradational stacking patterns or demonstrate progradational patterns. Lithologically, the MFS again shows evidence for an abrupt deepening over shallow peritidal micrites into much more heavily bioturbated wackestones facies that lack fenestral micrite cycle caps typical of the underlying TST.

In most cases MFS contacts occur at hardgrounds and show evidence for corrosion with pyritic or phosphatic coatings. Conodont skeletal lags are commonly developed at or near these intervals and K-bentonites appear to be more easily preserved and recognized in these settings. Hardground clasts or “platters” are often recognized at these contacts and are typically bored and occasionally encrusted by a number of different taxa. The presence of these clasts suggests that intermittent erosion can take place during development of the surface. This fact has suggested to some previous workers (i.e. Titus and Cameron, 1976, Joy et al., 2004) that these surfaces were subaerial unconformities. In fact the former authors placed the Black River-Trenton Group contact at such a surface in the Mohawk Valley, of New York. Nonetheless, these contacts show evidence for deepening and condensation suggestive of flooding. Hence, these contacts are interpreted by Brett and colleagues (2004) as surfaces of maximum sediment starvation, and associated with maximal rates of sea level rise and periods of non-deposition. Hence, the corrosion surface or interval at the top of the TST limestones may represent a considerable amount of time with little carbonate production and virtually no sediments recorded.

Early Highstand Systems Tracts HSTs

The Highstand Systems Tract (HST) is deposited after sea-level has reached its relative highest level and extends upward through the period where sea-level drops rapidly. These successions are characterized by aggradational to progradational small-scale cycle stacking patterns of small-scale cycles and are characterized by significantly more siliciclastic sediment than the underlying TST. In the Trenton, HST parasequences coarsen-upward and individual beds become thicker toward the top of the HST. This is typical of aggradational to progradational parasequence stacking.

McLaughlin and colleagues (2004) suggested that, especially for the Trenton sequences, HSTs were dominated by shaly nodular wacke-packstone and rhythmite facies. The input of siliciclastic mud occurs due to the first phases of sea-level drop which allows remobilization and introduction of estuary-sequestered sediments into offshore environments. In some cases, the input of siliciclastics and carbonates in HSTs appears to be rather rapid with the development of obrution-style deposits whereby fossil assemblages are essentially buried in situ. These likely reflect major events that may be related to climate shifts and/or oversized storms and reflect the down-slope transport of large volumes of sediments suspended from shallower water environments. In the pre-Trenton sequences, the broad platform architecture and relatively shallow-water environments prevented significant lateral transport of siliciclastic sediment under marine transport processes; thus, except in areas proximal to the Sevier Basin, successions remain relatively starved of siliciclastic sediments, although carbonate production is optimum as the carbonate factory is perfectly poised.

In the Trenton Group (Chatfieldian), proximal or shallowest water facies of the HST are characterized by no more than a few meters of wavy interbedded pack- to grainstones with thin argillaceous partings. In slightly deeper water environments down-ramp, they consist of interbedded wackestones, fine-grained calcisiltites, calcilutites, and calcareous shales. In the deepest or most distal environments within basinal settings of the Sebree Trough or the Taconic Foreland, HSTs are developed as dark gray to black, organic-rich shales with a relatively sparse fauna. Coincident with the upward increase in average grain size, as the HST develops, fauna generally become more abundant and robust in later portions of the HST. Taken together this suggests depositional environments were increasingly influenced by higher storm energies as sea-level fell (Holland et al., 2001).

Forced Regression Surface (FRS)

The term Forced Regression Surface (FRS) otherwise referred to as the "surface of forced regression," was used to designate the prominent surface developed within marine strata that demarcate the position or onset of major base-level fall (Hunt and Tucker, 1992; Catuneanu, 2006). Essentially sea-level change reaches its maximum rate of fall at this contact, and the shoreline is rapidly prograded basinward. The FRS thus separates the early highstand from the later highstand or regressive systems tract. Plint and Nummedal (2000) suggested that the FRS is another erosion surface that is marked by a facies dislocation. In many cases, in the Late Ordovician, this surface is represented as an abrupt shift from allodapic carbonate rhythmite succession of the early HST into shallower shaly nodular and/or calcarenite facies often with significant contributions of siliciclastic silt. The FRS can be subtle in basinal settings; however, it often becomes more pronounced in up-ramp settings. As suggested by McLaughlin and colleagues (2004), the FRS can display dramatic evidence for erosion and channeling into the uppermost beds of the underlying HST. In some cases several meters of HST can be removed in the proximal to medial portions of the ramp and in some cases, within the Lexington Limestone, much of the HST can be removed so that it is not easily recognized. Thus in up-dip or proximal settings, the FRS can be developed into the sequence boundary.

Late Highstand Systems Tracts or Falling Stage Systems Tract (FSST)

The FRS is typically sharply overlain by coarser skeletal wacke- to grainstone beds that also exhibit a shallowing-upward (or progradational) pattern. Shallowing is usually much more pronounced in this interval than in the underlying HST and the succession forms a basinward – thickening wedge of sediment between the FRS and the next sequence boundary surface. This interval has been referred to as a Regressive Systems Tract (RST) by Brett et al. (1990) or a Falling Stage Systems Tract (Plint & Nummedal, 2000). Brett and colleagues (2004) also used this term in the Late Ordovician as did McLaughlin (2007). In relatively proximal successions of post-Black River sequences, the FSST is often difficult to separate from the next overlying TST as strata of the FSST may at first approximation be similar in lithology to the sediments of the basal TST immediately above the SB. Alternatively, the FSST may be so thin, that it is unrecognizable unless evidence for multiple horizons of erosion or karstification can be identified (i.e. the FRS and the SB). However, careful examination of litho-, bio-, and taphofacies typically allows for recognition of the change in relative sea-level and the sudden reduction in siliciclastic sediments to be demonstrative of the change. The FSST is more easily recognized in down-ramp environments where it is thicker and becomes muddier and shows more pronounced channeling at its base. In these areas, the sudden change to coarser sediments helps signal the rapid progradation of shallow-water facies characteristic of the FSST over the finer-grained muddier facies of the HST.

Overall, the FSST represents at least a portion of the time during which deposition of the lowstand fan would occur below the shelf/slope break, as the majority of sediments forming the lowstand fan would be channeled through the FRS channels during development of an erosion surface in proximal regions. As suggested by McLaughlin et al. (2004), the term RST is likely the equivalent of the ramp margin wedge of Van Wagoner et al. (1990). Nonetheless, as the sequence boundary occurs immediately above the succession in question, the RST described by McLaughlin et al., is more closely aligned with the “falling stage systems tract” (after Plint and Nummedal, 2000).

REFINED SEQUENCE STRATIGRAPHIC FRAMEWORK

Recent sequence stratigraphic studies completed by the author and colleagues identified depositional sequences in the Black River-Trenton interval in the New York-Ontario outcrop belt (Cornell, 2001, Brett et al., 2004). These were tentatively linked to the sequences in the Nashville Dome and Cincinnati Arch region recognized by Pope and Read (1998) and Holland and Patzkowsky (1996, 1998). **Figure 6** shows a detailed comparison between the sequences of

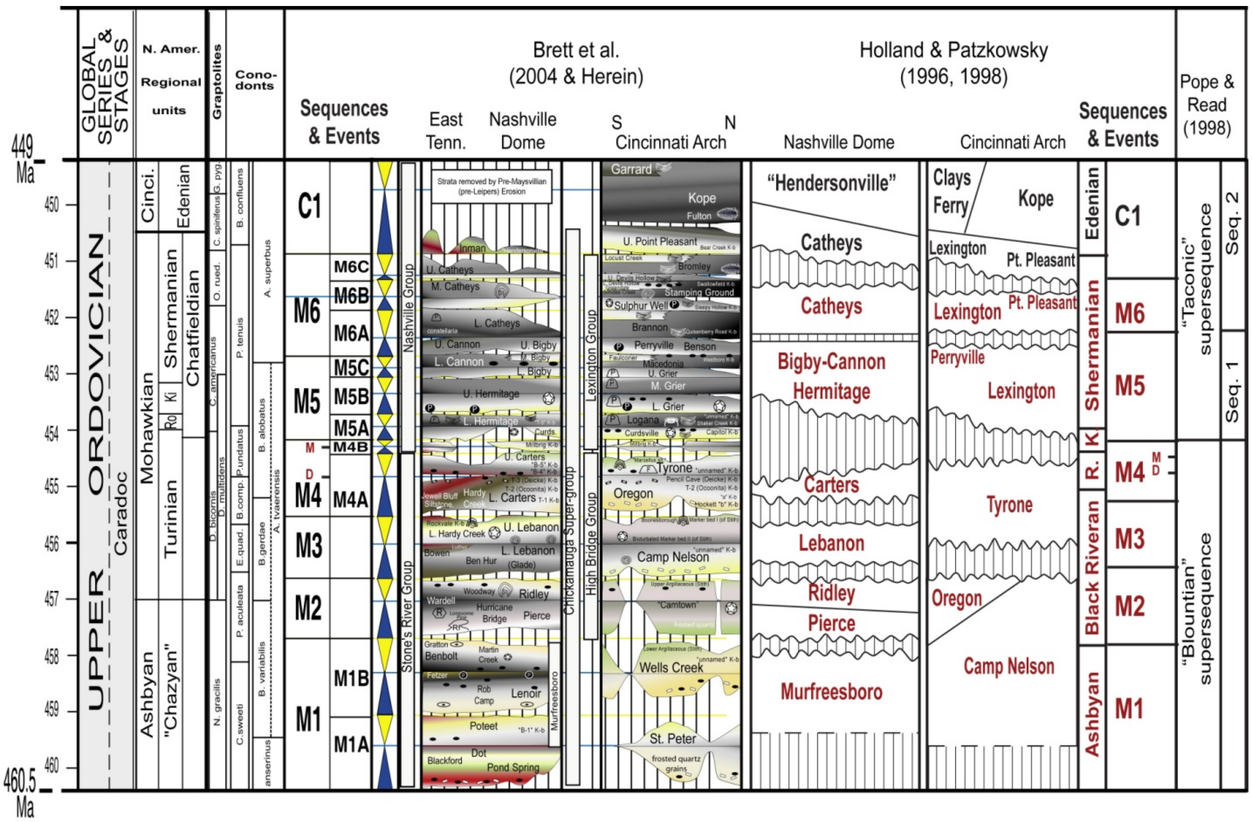


Figure 6: Detailed comparison between the sequences established by Holland & Patzkowsky (1996, 1998), those of Pope & Read (1998), with those established by Brett et al., (2004) and herein. Although the Nashville Dome successions are generally very similar (there are some minor modifications), the sequence framework of the Cincinnati Arch region is substantially modified from the original concept of Holland & Patzkowsky, both within the High Bridge Group and in the Lexington Limestone.

the Cincinnati Arch and the Nashville Dome compared to the sequences and stratigraphic nomenclature used by Holland and Patzkowsky (1998) for the same regions. As shown, detailed stratigraphic assessments in outcrop, and cores from the subsurface of the Cincinnati Arch region have enabled construction of a higher-resolution stratigraphic framework for this interval. As outlined elsewhere in this dissertation, numerous biostratigraphic, lithostratigraphic, and event stratigraphic correlations have enabled construction of a re-correlated stratigraphy. These have been supplemented here through recognition and correlation of sequence stratigraphic patterns and important surfaces that enable further refinement and a more robust chronostratigraphy for use in a more detailed analysis of the GACB during this time. From this initial work a more comprehensive inter-regional set of third-order sequences has been constructed here for the entire interval from the base of the Chazy (earliest Ashbyan) to the top of the Mohawkian (latest Chatfieldian) for the Cincinnati Arch, New York-Ontario Platform, and the Pennsylvania Ridge and Valley Province (**figure 7**).

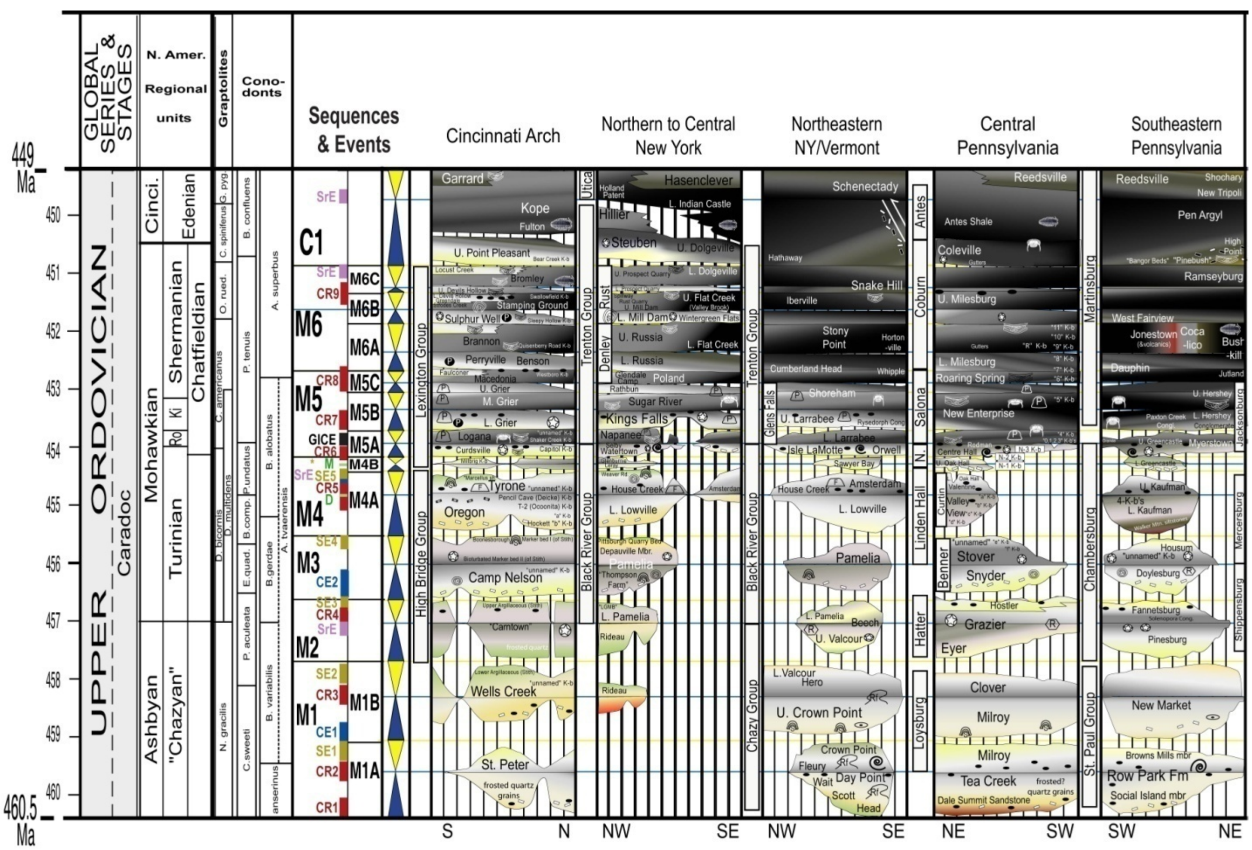


Figure 7: Depositional sequences for the Cincinnati Arch, New York-Ontario Platform, and central and eastern Ridge and Valley Province of Pennsylvania. Also shown are key biostratigraphic zones and events recognized through detailed investigations discussed elsewhere. Event abbreviations: CR1: Chert-Rich Interval 1, SE1: Siliciclastic Event 1, CE1: Calcification Event 1, SrE: Strontium Isotopic Excursion, M: Millbrig K-bentonite, D: Deicke K-bentonite, GICE: Guttenburg Isotopic Carbon Excursion. (see Appendix A2 for larger figure)

M1-M4 SEQUENCES OF THE "BLOUNTIAN SUPERSEQUENCE"

The M1-M4 third-order sequences of Holland and Patzkowsky (1996, 1998) collectively record the maximum extent of the GACB during the Late Ordovician, coincident with the onset of the first or Blountian Tectophase of the Taconic Orogeny. Thus here Pope and Read's (1998) "Blountian Supersequence", is subdivided into five relatively thick (10s of meters) sequences (M1A, M1B, M2, M3, M4A) and a sixth relatively thin, but distinct "mini-sequence" at the top of the M4 sequence (termed M4B). These carbonate-dominated sequences overall show relatively shallow sub-tidal to peritidal deposition where water depth was likely never deeper than a few meters and certainly never below the photic zone in the study areas discussed here. These sequences show numerous well-developed small, meter-scale, cycles that generally can be correlated in closely-spaced outcrop sections throughout local outcrop areas (see Cornell, 2001). To date, it has not yet been possible to correlate these cycles between different outcrop regions separated by more than 200 km. Nonetheless, unique facies including chert-rich intervals, siliciclastic-rich intervals, K-bentonite horizons, chemostratigraphic events, etc. do allow for inter-regional correlation. The details of depositional sequences in this interval are outlined below, although detailed descriptions of lithologies and sequence architecture are discussed elsewhere herein (and see Brett et al., 2004; McLaughlin et al., 2004; McLaughlin, 2007).

M1A Sequence: St. Peter –Day Point-Tea Creek-Row Park

SB & TST: TYPE CHAZY REGION

The lowest sequence in the study interval (**figure 8**) begins immediately above the Knox Unconformity which in northeastern New York and western Vermont separates the Beekmantown Group below from the Chazy Group above. Specific details of the basal sequence boundary in New York are discussed in detail in chapter three of this dissertation. In general the

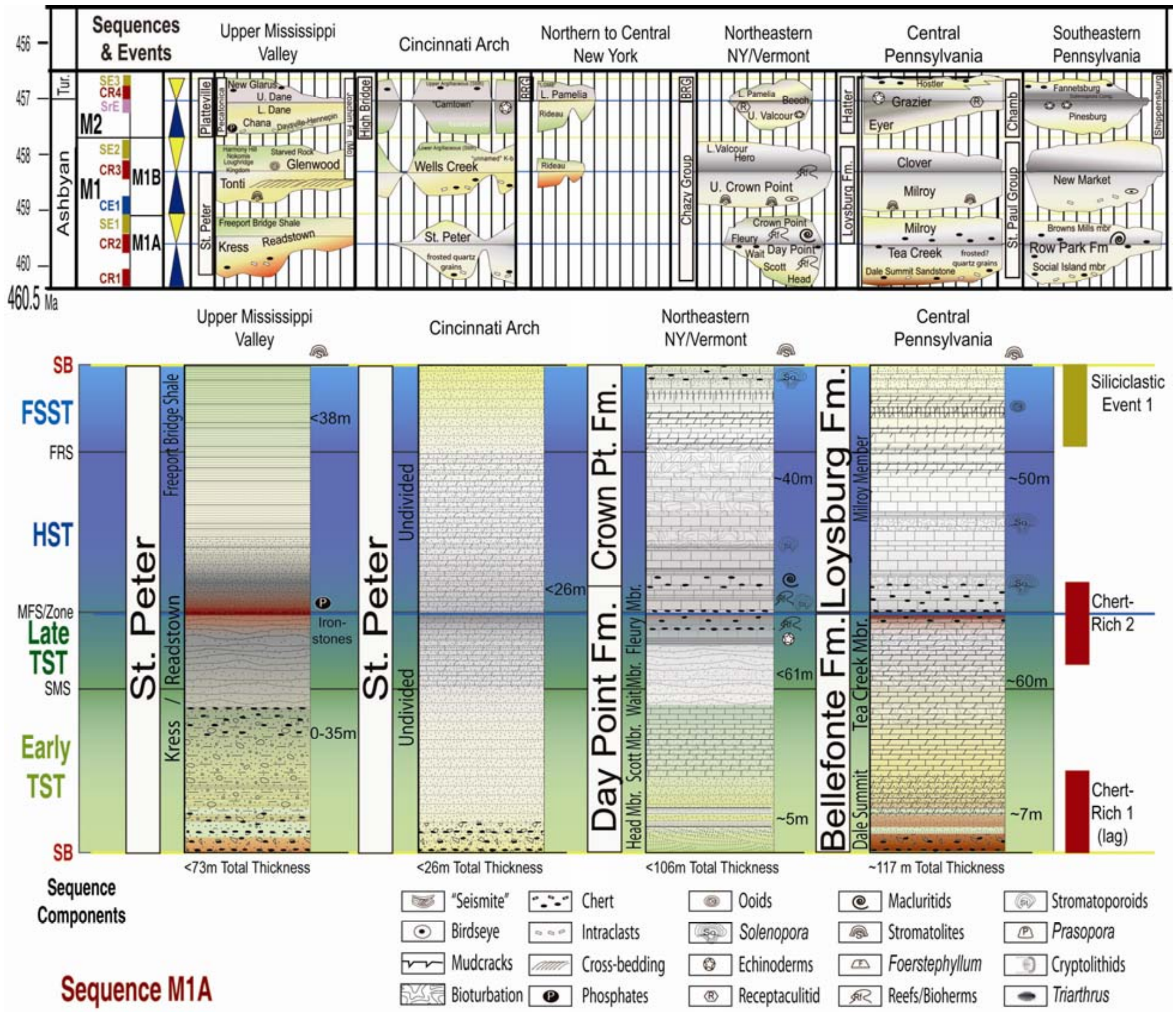


Figure 8: Correlated framework of the M1A depositional sequence (Ashbyan) for the Upper Mississippi Valley, Cincinnati Arch, Northeastern New York, and central Pennsylvania. Sequence is thinnest in the vicinity of the Jessamine Dome/Cincinnati Arch area where the sequence is occasionally not present, and thickens both northwestward (where it is dominated by coarse-grained facies of the lower St. Peter Group), and eastward where the unit becomes dominated by carbonates (central Pennsylvania) and mixed siliciclastic/carbonate facies (New York).

sequence boundary juxtaposes mature quartz sandstones interbedded with occasional polymictic conglomerates onto the Beekmantown Group carbonates. This horizon may contain small clasts of reworked cherty limestones and represents chert-rich interval 1 discussed in chapter 6.

The lowermost unit of the Chazy Group is the Day Point Formation which is composed of four members: the Head, Scott, Wait and Fleury Members. Although exposures are generally poor and a complete section is lacking in the Champlain Valley, descriptions and assessments of the Day Point indicate that it is a heterolithic unit. Internal small, meter-scale parasequences may show dramatic variation laterally owing to irregular topography of the region as afforded by the irregular karstic sequence boundary below and suspected local faulting. Falkenberg and Mehrtens (1993) recognized repeating cycles containing at least seven different lithofacies, which include: 1) highly bioturbated sandstones; 2) bioturbated, brachiopod-rich wackestones; 3) planar cross-bedded, interbedded sandstones and sandy packstones; 4) interbedded sandstones and shales grading into thinly interbedded sandstones and limestones; 5) grainstones with mound-shaped bryozoan “reefs” or thin, non-reefal, layers of bryozoans and planar laminated and cross-stratified beds; 6) fine-grained calcareous sandstones with planar and herringbone cross-stratification; and 7) planar cross-bedded packstones with bryozoan reefs or thin non-reefal layers of bryozoa. As also noted by other authors, Falkenberg and Mehrtens indicate that there is substantially more sandstone in the lower portion of the Day Point, and it is overlain by deeper water carbonates.

Shaw (1969) previously suggested the interval represented an overall deepening-upward pattern. Thus, this basal unit is interpreted as a TST deposit with only the uppermost part of the Fleury Member occurring above the MFS. In New York and the western Lake Champlain Islands, the Day Point is dominated in its lower half by quartz sandstones, calcareous sandstones, and sandy limestones that are arranged into shallowing-upward cycles that become progressively more fossiliferous upward with occasional argillaceous partings weathering in slight recess to rubbly textures. The upper portion of the formation is dominated by somewhat deeper-water, fossiliferous wackestones and grainstones containing numerous biostromes (Oxley & Kay, 1959). To the east, into Vermont, Oxley and Kay (1959) show that the Head Member grades laterally into much finer-grained facies dominated by fine-grained, calcareous quartz sandstone to greenish (glauconitic?) siltstones interbedded with silty argillaceous seams. This unit is called the St. Therese Siltstone and it disconformably overlies the Beekmantown Group. It is likely the down-ramp equivalent of the coarser-grained strata to the west. At the south end of Lake Champlain in the Crown Point type-section, the basal members of the Crown Point are not developed and only the middle to upper Crown Point Members appear as they overstep and onlap the sequence boundary toward the Adirondack Arch and the Mohawk Valley.

The uppermost unit of the Day Point Formation, the Fleury Member, is here interpreted to contain the maximum flooding zone and is the most widespread and most uniform lithologically. The Fleury is typically a two part unit. The lower part of the Fleury is composed of light-gray, highly fossiliferous, brachiopod-bearing, cross-bedded grainstones containing small biohermal masses that represent the typical “Crown Point lithology.” In Vermont a particularly prominent interval near the top of the lower Fleury (between 13-15 meters above the base) contains a zone of bioherms with the earliest tabulate corals *Lamottia heroensis* (Oxley & Kay, 1959). Representing some of the earliest coral “reefs” (Raymond, 1923), the bioherms are generally less than six meters in diameter and may have had up to one meter of relief (Fisher, 1968). These bioherms appear to have developed on hardground surfaces at the top of at least two small-scale shallowing upward cycles. This interval, although rarely exposed today, was quarried for decorative building stone as the “Lepanto Marble” in the Plattsburg region of New York (Fisher, 1968), because it carried a distinct red, iron staining typical of some condensed hardgrounds. The top of the *Lamottia* bioherm zone is demarcated by a sharp facies dislocation to much-finer grained grainstones (calcisiltites as per Oxley and Kay, 1959) interbedded with minor shales that demarcate the upper sub-unit of the Fleury. As suggested by Oxley and Kay, this upper Fleury is difficult to separate lithologically from the lower Crown Point facies that overlies it and is interpreted here to be the lower part of the HST.

MFS-HST OF TYPE CHAZY REGION

The facies dislocation surface described above is interpreted as the MFS of the M1A sequence in the type Chazy region. Although detailed correlation of this interval throughout the Champlain Valley is still necessary, it is clear that the Fleury is the most widespread unit of the Day Point. The maximum flooding zone shows an abundance of nodular to slightly bedded cherts in some localities. In some cases bryozoan zooecia are infilled with chert or are in some cases silicified. This chert-rich interval is CR-2 (**see figure 7**). As mentioned in chapter 3, the upper Fleury contains multiple reddish, iron-stained hardgrounds that likely record the condensed interval.

The HST of the M1A sequence is represented by the uppermost Fleury sub-unit and the lower Crown Point Formation. The upper Fleury is dominated by small-scale cycles of thin to massive, bioturbated calcisiltites and fine-grained calcarenites interbedded with wackestones and packstones that typically are separated by shaly-nodular horizons. This fine-grained calcarenite interval is approximately twenty meters-thick, is typically fossiliferous and shows occasional zones of normally graded packstones to grainstones resting on large bryozoan reefs or thin planar, cross-bedded layers composed of fragmented bryozoa and occasional *Maclurites magnus*. These facies appear to record numerous storm-deposits interbedded with lower-energy calcisiltites deposited below normal wave base. Thus the upper Fleury likely represents the deepest facies of the M1A sequence. About six meters from the top of the unit, large numbers of *Stromatocerium* and bryozoans occur in biohermal masses that become characteristic of the uppermost Fleury.

Like the underlying upper Fleury, the lower part of the Crown Point is described as an argillaceous medium to fine-grained, even sub-lithographic, bioturbated limestone (Welby, 1961). As with the upper part of the Fleury, the Crown Point also is noted for its *Stromatocerium* bioherms that are more diverse in that they contain another early stromatoporoid referred to as *Cystostroma vermontense* as well as early rugose corals. Unlike Day Point bioherms, these are more diverse and all lack the coarse-grained flanking calcarenites, which was interpreted by Shaw to indicate that these facies were deeper than the underlying Day Point.

Nonetheless, near the top of the lower Crown Point, Shaw (1969) noticed the appearance of thin to very thin (1-2 cm-thick) ribbon beds of dolomitic grainstones to silty lime mudstones containing occasional quartz grains. In some cases, coarser quartz-rich materials are piped down into vertical *Phytopsis*-style burrows. The quartz-rich intervals become more predominant in western and southernmost outcrop areas close to the Adirondack Arch/Beauharnois Arch and become more pronounced upward into the middle of the Crown Point Formation. Quartz grains are also noted to form the cores of small centimeter-sized *Girvanella* algal nodules. This upward change to siliciclastic-influenced, ribbon bedding (almost pin-striped in some cases) is typical of late HST (or FSST) facies documented in the Lexington Limestone by McLaughlin and colleagues (2004). Here this suggests rapid shallowing and progradation of nearshore quartz-rich facies over the finer-grained carbonates during the latest HST.

The top of the M1A sequence is not yet clearly defined, but the upper Crown Point was described by Griffing (2000) to contain a number of domal to columnar stromatolites that were interpreted as late-stage components of biostrome communities. Griffing showed these units to contain *Solenopora* red algal crusts, birdseye limestones, mudcracks, and disconformity surfaces developed on top of and in channels between reefal facies during facies restriction events. Mehrtens (1998) suggested that the channels were generally oriented perpendicular to oncoming wave fronts (and likely developed as spur and groove structures). Here it is suggested that these facies might be interpreted as indicative of the latest stages of regression and/or the very earliest phases of the subsequent sea-level rise. These channels may have formed during the rapid lowering of sea-level during the regressive systems tract, and potentially filled during the ensuing transgressive systems tract. If this is the case, the channelized interval likely represents a composited FRS and SB surface, and in contrast to Griffing (2000), the stromatolite-bearing facies may have been developed as a pioneer community on the relict topography rather than a climax community.

CORRELATION OF THE M1A SEQUENCE

The M1A sequence, the lowest sequence of the type-Chazy, is not recognized in the New York-Ontario outcrop region outside of the Champlain Valley. Its development was likely restricted in this region by the development of the Beauharnois Arch and the Adirondack Arch at about this time (Salad Hersi et al., 1998). Nonetheless, outside of the New York-Ontario outcrop region and the New York promontory, it appears that age equivalent sequences were deposited. Here, using biostratigraphy and event intervals as discussed, the M1A sequence is correlated with the lower part of the St. Peter Sandstone of the Upper Mississippi Valley and the Cincinnati Arch (**see figure 8**).

Upper Mississippi Valley

In the Upper Mississippi Valley, the TST of this sequence is represented by the Readstown or Kress Members of the St. Peter Formation. Ostrom (1967) identified the Readstown Member in Wisconsin as the basal conglomerate of the St. Peter Sandstone. It is equivalent to the Kress Member identified by Buschbach (1964) in the subsurface of northeastern Illinois. The unit consists of a few meters of breccia composed of sandstone pebbles, cobbles, and even larger boulders embedded in a sandy matrix containing silt and clay. The unit is irregularly bedded and grades upward into a zone of nearly black ironstones that show evidence of dark-red iron oxide coatings (Lee & Attig, 1990), thus suggesting transgressive condensation. In Wisconsin, the Readstown sits above the Oneota Dolostone Formation of the Prairie du Chien Group, which is Early Ordovician in age. In some places, the St. Peter sits on strata of the Upper Cambrian (Dott et al., 1986) suggesting a significant amount of topographic irregularity and incision into underlying units. Thus, the contact at the base of the St. Peter is the Knox Unconformity and is the sequence boundary of the M1A sequence. Witzke and Bunker (1996) suggested this portion of the M1A showed evidence for onlap of older relict topography, but defined the unit to represent one short-term transgressive-regressive cycle in Iowa (Cycle A of their broader scale “Ancell sub-cycle”).

The HST of the M1A sequence in this region is represented by the newly recognized “Freeport Bridge Shale” (Young et al., 2005), which is identified herein as representing siliciclastic event 1 (see chapter 6). In northeastern Iowa to the west of the Wisconsin Dome, these authors recognized a distinct laminated greenish to dark gray shale immediately below the Tonti Member of the St. Peter and above the lowermost sandstones of the St. Peter. The shale unit appears to contain a lagerstätte deposit containing well-preserved faunas including Chazy-aged conodonts, macrophytic algae, and crustaceans (Liu et al., 2006). As it sits immediately above the basal St. Peter and contains a marine (albeit restricted) fauna, it is interpreted as the HST of the M1A sequence. This interval corresponds to T-R cycle B of the “Ancell sub-cycle” as defined by Witzke and Bunker (1996).

In its only known outcrop exposure, the Freeport Bridge is shown to range from nearly 38 meters down to less than three meters where it appears to be truncated above by the upper St.

Peter. The Freeport Bridge may have been deposited in an embayment or estuarine area that was subsequently eroded and or infilled during the later HST progradation of the St. Peter or during the ensuing transgression. Where sandstones are more prominent, and the Freeport Bridge is not present, Witzke and Bunker indicate the occurrence of a phosphatic to iron-stained, organic-rich shale that likely marks the approximate position of the Freeport Bridge shale. It likely represents the MFS interval of the sequence.

Cincinnati Arch

The M1A sequence is not exposed in the Cincinnati Arch region. It also does not appear to be represented across the Jessamine Dome or along the axis of the Arch into northern Kentucky, except for in a few small pockets. Nonetheless, the St. Peter has been recognized in the subsurface north of the Cincinnati region, on the southeastern flank of the arch, and on the western flank of the arch where it reaches its maximum thickness of up to 43 meters and is included in the base of the Ancell Group (Droste et al., 1982). The overlying Wells Creek and the St. Peter are known to contain Ashbyan conodonts (Sweet & Bergström, 1976; Droste et al., 1982). It is not well-studied and details are minimal. However, the base of the unit has long been equated with the Knox unconformity and the St. Peter itself appears to onlap localized uplifts in central Kentucky through northern Kentucky where the unit is usually very thin or not present. It does not appear that the Cincinnati Arch itself was a pronounced feature, but it appears that at least some localized areas along the present day arch were uplifted and impacted deposition during the M1A sequence.

In southwestern Ohio and the northern Jessamine Dome region, Kentucky, Calvert (1962) and Carpenter (1965) recognized significant and pronounced erosional topography on the Knox Dolomite Group. In this area, this topography results in the variable thickness of the St. Peter and thus the St. Peter here is interpreted to represent the entire M1A sequence. Here the St. Peter, although still a sandstone, is typically more calcareous and contains occasional dolostone beds within it. Carpenter (1965) noted that many of the quartz grains were well-rounded and even frosted suggesting they may have been transported at least in part by wind. To date it is not clear if there is a Freeport Bridge equivalent and therefore it is difficult to distinguish further details of this sequence. Nonetheless, the overlying Wells Creek (here considered to represent the M1B sequence) helps constrain the M1A sequence as discussed below.

Central Pennsylvania

In central Pennsylvania, the basal sequence boundary of the first Chazy sequence appears to coincide with the Knox Unconformity that sits below the top of the Bellefonte Dolostone. In central Pennsylvania the gap at the base of the M1A sequence is difficult to establish owing to the heavily dolomitized nature of the Bellefonte; however, it is likely similar in duration to that recognized in the type-Chazy region by (Landing & Westrop, 2006). In contrast to the Upper Mississippi Valley, where the M1A sequence is siliciclastic dominated, the M1A sequence is dominated by carbonates, which is also the case in the Champlain region. The two units at the top of the Bellefonte (the Dale Summit Sandstone and the Tea Creek Dolostone of Swartz, 1955) along with the overlying lower Milroy Member of the Loysburg Formation are recognized here as the M1A sequence. In total, the sequence ranges to about 150 meters in the northwestern Nittany Valley near Bellefonte. The occurrence of occasional lithoclasts and well-rounded quartz grains in the Dale Summit are suggestive of the first supra-Knox sequence elsewhere. Moreover, correlation of the Milroy with the upper Day Point Formation of New York using biostratigraphy and the occurrence of chert-rich interval 2, enables this sequence to be relatively well-constrained. As suggested here, the relatively restricted facies of the Dale Summit to Tea

Creek Dolostone appears to represent the TST. The Tea Creek consists of massive fine-grained dolostones that contain some evidence of bioturbation but contains no other recognizable faunal elements. The unit, however, does contain an abundance of iron-stained contacts and reddish mottling near the top of the unit. This interval is thus thought to represent poorly preserved mineralized hardground intervals formed during the late stage of the TST.

As mentioned in chapter 4, the traditional base of the Chazy has been placed at the top of the Tea Creek Member, but this was contested as the contact is typically planar and sharp rather than irregular and erosive. Wagner (1963) suggested this contact was not a substantial unconformity and likely represented a minor shift in lithology. Nonetheless, the contact described by Wagner is interpreted as an MFS surface that juxtaposes the limestone-dominated facies of the Milroy Member of the Loysburg Group above the reddish-stained dolostones of the Tea Creek.

The Milroy Limestone is typically a medium to dark gray, “calcisiltite” (very fine-grained grainstone) interbedded with thin (2-4 cm-thick) light-gray to buff weathering dolomitic limestones. This unit characteristically weathers as a “ribbon” limestone or the “Tiger-Stripe” by Kay (1944). In the lower to middle portion of the Milroy, Swain (1957) recognized some minor algal (*Solenopora*?) bioclastic breccias and a slightly higher zone containing a few ooids and domal stromatolites associated with ostracod and bumastid trilobite-bearing packstone beds capped again by another algal breccia. As described, this facies and the observed pattern is very similar to that described for portions of the upper Day Point (Fleury Member) to lower Crown Point Formation of the Champlain region and represent a shallowing upward pattern. The occurrence of the stromatolites in the middle Milroy is likely synchronous with those recognized by Griffing (2000) in the Champlain region in the Crown Point. This succession thus is interpreted as the equivalent of the M1A HST to SB interval of the M1A to M1B sequences.

M1B Sequence: Wells Creek-Crown Point-Milroy-New Market

SB-TST OF TYPE CHAZY REGION

The next sequence of the Blountian Supersequence is here termed the M1B sequence (**figure 9**). It is reflected in the Champlain Valley in strata of the upper Crown Point Formation through the overlying lower Valcour Formation. Although the Crown Point has not previously been divided formally into members, the overlying Valcour has. The latter has been subdivided into two units, the Hero and Beech Members. The Hero Member is here combined with strata included in the upper Crown Point Formation to form the M1B sequence, while the Beech Member forms the TST of overlying sequence M2 (see below). Overall the Valcour is the most argillaceous of the Chazy Group; this clastic influx is coincident with the onset of major shale deposition in the Sevier Basin further south and represents siliciclastic event 2 as discussed in chapter 6.

As suggested by Griffing (2000), strata formerly assigned to the uppermost Crown Point Formation contain evidence of restricted, shallow-water deposition. The unit includes the first fenestral micrites, increased abundance of algal facies, and quartz sandstones deposited in channels on top of deeper-water Crown Point reefs. As mentioned, the sharp juxtaposition of shallow-water facies over the deeper-water reefs conforms to the pattern of a sequence bounding interval. The appearance and progradation of quartz-bearing ribbon limestone facies (siliciclastic-event 1) followed by thin quartz sandstones interbedded with fine-grained, dolomitic limestones, and even mud-cracked, birdseye facies with domal stromatolites suggests

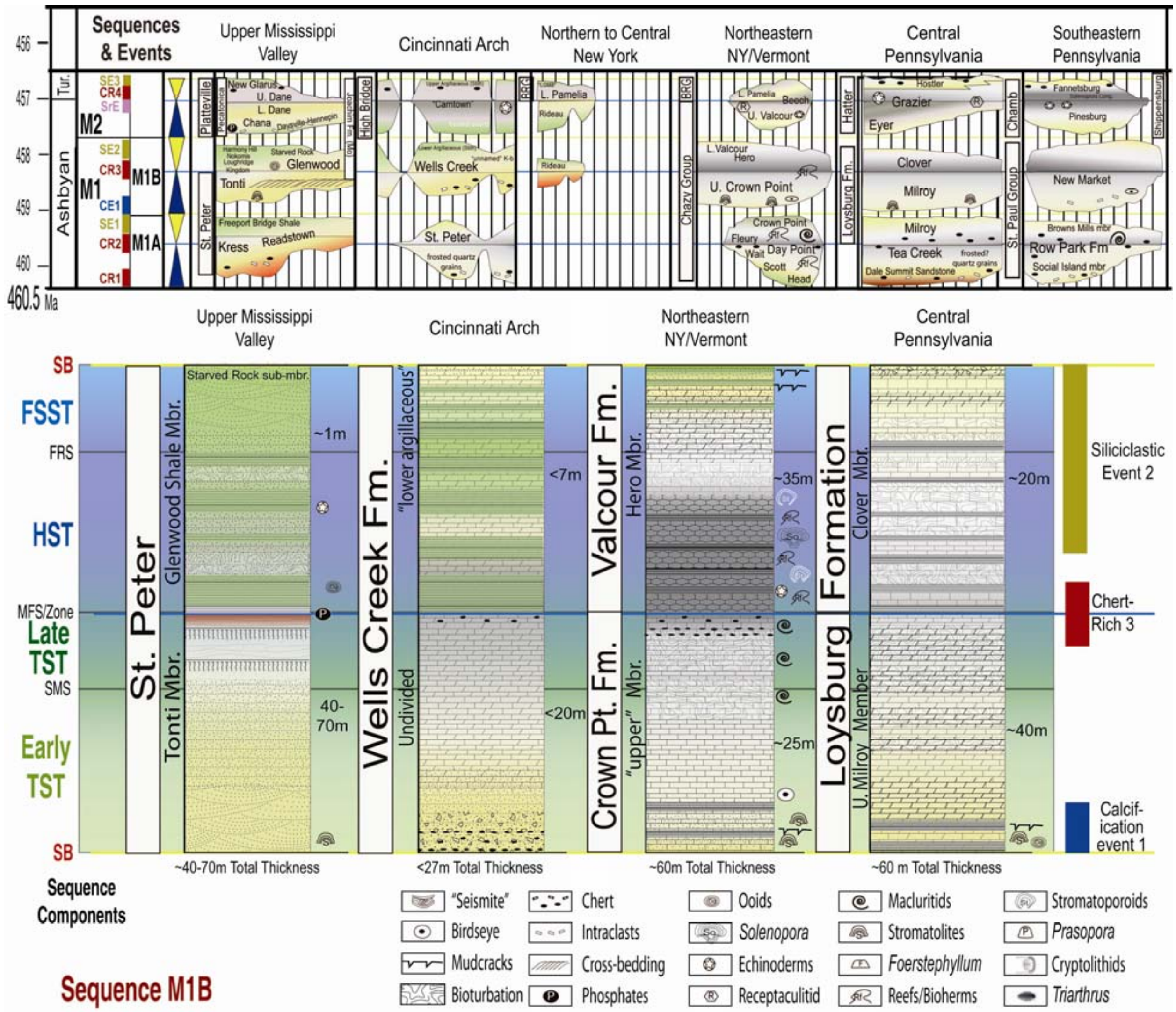


Figure 9: Correlated framework of the M1B depositional sequence (Ashbyan) for the Upper Mississippi Valley, Cincinnati Arch, Northeastern New York, and central Pennsylvania.

initiation of a new TST. These fine-grained units are succeeded upward by slightly coarser-grained carbonates that lack coarse siliciclastics and are again, for a short-time, relatively clean and devoid of siliciclastics indicating that the supply of siliciclastics is again reduced as a result of base-level rise and sequestration. These uppermost Crown Point facies also contain abundant cephalopods and gastropods including large numbers of *Maclurites magnus.* The latter are abundant on some bedding planes in the upper Crown Point where they appear to be reworked occasionally and condensed into planar pavements. Here this succession is interpreted as an upward-deepening pattern typical of a TST.

MFS - HST OF TYPE CHAZY REGION

The Hero Member of the Valcour Formation in Vermont consists of dark gray fine-grained calcarenites and argillaceous calcisiltites arranged into meter-scale cycles that often weather to shaly nodular rubble, but rarely show evidence of deepening or shallowing overall and appear aggradational. The shaly-nodular bases of these cycles may contain well-preserved fossils including at least one species of unidentified cystoid and is overall very similar in lithology and appearance to the Benbolt Formation of the southern Appalachians. In New York, the Hero is a deeper facies than found in the uppermost Crown Point below. It is likely that the contact of the Hero with the underlying Crown Point is the MFS or at least contains the maximum flooding zone, but due to poor exposures this contact is in need of significantly more study.

The HST of the M1B sequence is interpreted to be represented by the Hero Member of the Valcour Formation. As mentioned in chapter 3, the Hero contains the third prominent interval of reefs in the Chazy region. Like most Chazy reefs, these are constructed by bryozoans, sponges, algae, and stromatoporoids. These reefs contain abundant *Solenopora,* minor *Stromatocerium* and numerous specimens of the lithistid demosponge *Zittelella varians* (Oxley & Kay, 1959), the latter of which is also found in the Murfreesboro and Ridley Limestones of Tennessee (Wilson, 1949). Reefs in this interval eventually grade upward into slightly coarser facies and become interbedded with medium-light grey calcarenites indicating shallowing. There is also evidence for channeling and a return to tidally-influenced environments indicated by the occurrence of herringbone cross-stratification. These channels are commonly filled with trilobite and nautiloid fragments, carbonate sands, and occasional siltstone beds. The succession was interpreted as a shallowing by Shaw (1969). This interval represents the second progradation and channel-filling succession of the type-Chazy region and the top of the M1B sequence.

To the west of Lake Champlain, at Chazy, New York, the Hero Member is mostly composed of thin interbedded dolostones and calcilutites separated by minor siltstones and argillaceous partings. In this region, in board of the main reef-bearing facies, the Hero Member is dominated by another ribbon facies and appears to lack the reefal facies. Dolostones are more predominant where the beds were intensely bioturbated and likely represented lagoonal facies. These beds commonly show evidence for restricted faunas, greenish-gray quartz siltstones and minor sandstones (representing siliciclastic event 2). Here this succession is interpreted as the final progradational interval of the late HST ranging into the sequence bounding interval at the base of sequence M2. Clearly more work is needed to refine the position of the M1A – M2 sequence boundary. However, the loss of substantial siliciclastic beds, the presence of desiccation features, and the return of minor cryptalgal lamination in the dolostones, and minor fenestral micrites of the overlying Beech Member suggest the initiation of yet another cleaning-upward pattern associated with siliciclastic sequestration of the next sea-level rise event leading into the M2 sequence.

CORRELATION OF THE M1B SEQUENCE

The M1B sequence is the first widespread and persistent depositional sequence of the Chazy Group in the type New York-Ontario region. Its equivalents are known along the St. Lawrence River and eastward. To the west of the Lake Champlain region, it is discontinuous across the Beauharnois Arch through the west side of the Frontenac Arch. In Ontario, the M1B sequence, likely correlates with portions of the Shadow Lake Formation of the Lake Simcoe

Region, and the Rideau Sandstone of the Kingston area of Ontario. Outside of these areas, the M1B sequence appears to correlate across the Michigan Basin, where depositional sequences were defined by Nadon and colleagues (2000). In the Upper Mississippi Valley region the sequence is represented by the upper St. Peter Sandstone and the overlying Glenwood Formation.

Upper Mississippi Valley

The SB interval of the M1B sequence is interpreted to occur at the contact between the Freeport Bridge shale and the overlying Tonti Member of the St. Peter Sandstone. Where the underlying shale is truncated, the Tonti Member rests on the Readstown/Kress conglomerate Members and the contact between them is generally placed at the highest conglomerate zone (Dott et al., 1986). The succession appears as a fining-upward succession, and a distinctive contact is not usually identified. Nonetheless, Templeton and Willman (1963) indicate that it is the most widely distributed member of the St. Peter and is referred to as a “sheet-sand” deposit by Dott and colleagues (1986). The Tonti Member consists primarily of poorly-cemented, fine-grained, well-sorted, non-calcareous quartz sandstone although some medium- to coarse-grained beds do occur in areas thought to have been deposited under fluvial influence.

The Tonti has been well-studied and has been shown to have been influenced by fluvial transport of sands as indicated by trough cross-bedding in some locations, especially near the Wisconsin Arch. Significant aeolian transport and large-scale dune development followed deposition of the fluvial successions (Dott et al., 1986). The aeolian deposits, in turn, give way upward to finer-grained and significantly more bioturbated facies including *Skolithos* burrows. Initially, the Tonti quartz-sands appear to have infilled relict incised channels during initial sea-level rise. Subsequent subaerial migration of dunes across the platform filled in any remaining topographic lows. The dunes were, in turn, beveled and reworked during marine transgression.

In terms of sequence stratigraphy, the lowermost Tonti behaves somewhat like a RST deposit with the truncation of underlying Freeport Bridge shale and formation of channels possibly representing FRS development. However, the infilling of channels followed by regional beveling, likely represents the redistribution of sediments during formation of the SB. Thus, it remains to be seen if some of the Tonti sits below the SB and/or if the SB is superimposed on the FRS, but it is likely that much of the sandstone was actually deposited during the LST to TST phase, and therefore above the SB of the M1B sequence.

In Wisconsin, only the uppermost Tonti appears to have been deposited under marine influence and it demonstrates a range of traces including vertical *Skolithos* burrows, and rare fossil debris. Many of these marine-influenced beds appear to the south and west of the Wisconsin Arch where planar beds were thought to have been stabilized by cyanobacterial mats (Dott, 2003). In this direction, the entire Tonti appears to have been deposited under marine influence. Thus, the Tonti shows evidence of transgression, and is sharply overlain by the marine-influenced Glenwood Formation. The Tonti is thus considered the TST of the M1B sequence. This is in slight contrast to the interpretation of Witzke and Bunker, who considered this portion of the St. Peter to represent the third (cycle C) transgressive-regressive sub-cycle of the Ancell Group.

The MFS to HST interval of the M1B sequence is represented primarily by the Glenwood Formation. The Glenwood is typically green shale to argillaceous sandstone interpreted historically as a marine shelf shale and/or lagoonal shale (see Ostrom, 1964; Fraser, 1976). In Wisconsin, the Glenwood is a thin, intensely bioturbated argillaceous sandstone to shale unit. At its base is a coarse-grained lag composed of phosphate nodules, and mineralized discontinuity surfaces. It is overlain by no more than a meter of green sandy shales. It was interpreted by Schutter (1996) as a condensed section typical of the latest transgression to early highstand. This entire unit was considered to be the uppermost of the transgressive –regressive cycles (cycle D) of the broader-scale Ancell Group sequence (of Witzke and Bunker, 1996).

Although fossils are typically not common in most of the unit, conodonts, scolecodonts, and chitinozoans are abundant in the lower portion of the unit (Schutter, 1996) and provide a Chazyan age. In Minnesota, the Glenwood contains at least one white-weathering clay layer that was interpreted to be a possible volcanic ash. Thus far this has not been confirmed (see Chetel et al., 2005), but given recognition of this interval as siliciclastic event 2, the coincidence of a suspected K-bentonite in the Sevier Basin in the Blockhouse Shale at this approximate level might represent the first K-bentonite deposited during the Blountian Tectophase of the Taconic Orogeny.

The upper Glenwood, near the contact with the overlying Platteville Limestone, contains trilobites, echinoderm fragments, bryozoans, and ostracodes and occasional pyritized sponge spicules and fecal pellets (Schutter, 1996). This faunal succession is clearly a marine influenced facies and likely represents the HST of the M1B sequence. As shown by Witzke and Bunker (1996) the uppermost sandy interval of the Glenwood (the Starved Rock Sandstone Member) progrades over the top of the shale. This sandstone thus represents the later HST facies, although it does not develop to nearly the same extent as the underlying St. Peter and is immediately capped by the Pecatonica Formation of the Platteville Group. Above this position, siliciclastics are severely restricted during the spread of the Platteville carbonates near the end of the Ashbyan during sequence M2.

Cincinnati Arch

In the Cincinnati Arch region, the M1B sequence corresponds to the Wells Creek Formation. Details of the Wells Creek stratigraphy are outlined in chapter 5. The Wells Creek has been correlated with the Glenwood Shale of the Upper Mississippi Valley for some time (Wickstrom et al., 1992) and at its base are equivalents of the upper St. Peter. In the subsurface of the Jessamine Dome region, it is an interval dominated by green, argillaceous dolostones and tan to green sandy dolostones below the lowest lithographic limestones of the High Bridge Group. The top of the Wells Creek contains an interval of interbedded dark gray to green shales (Gooding, 1992).

In northern Kentucky and southwestern Ohio, the Wells Creek is characterized by dolostones with a conglomeratic base containing thin lenses of sandstone which grade up into finer grained, greenish-gray dolostones containing at least one possible K-bentonite (Kolata et al., 1996). As elsewhere, this succession suggests that the base of the Wells Creek is unconformable and represents a SB. Carbonate-dominated facies of the ensuing TST grade upward into late HST deposits characterized by the finer-grained dark gray to green shales and interbedded dolostones of the upper portion of the Wells Creek. As in New York and in Pennsylvania, the top of the Wells Creek appears to grade between ribbon facies and thin beds of green shale containing thin dolomitic limestones similar to the Glenwood Shale of the Illinois Basin (Carpenter, 1965). This is the “lower argillaceous unit” of Stith (1979) and is interpreted here as representative of siliciclastic event 2. Further north in Ohio, Wickstrom and colleagues (1992) note that these beds to grade into thin brown, gray and black shales interbedded with thin sandstones and siltstones. These appear to prograde over the dolostones of the lower Wells Creek TST. Here the upper Wells Creek is interpreted to represent the latest HST interval of the

M1B sequence. The upper Wells Creek beds contain minor frosted quartz grains suggestive of the St. Peter, and may be reworked locally from the St. Peter in the vicinity of topographic highs during the late M1B – early M2 sequence interval. As this interval is entirely recorded in the subsurface, identification of the specific position of the SB at the top of the M1B sequence is difficult. However, these siliciclastic-influenced beds grade upward into relatively clean micritic limestones and dolostones of the lower High Bridge Group that indicate another major rise in base-level associated with the M2 TST.

Central Pennsylvania

Sequence M1B in central Pennsylvania is composed of the upper Milroy and the Clover Members of the Loysburg Formation. As mentioned, in chapter 4, Swain (1957) divided the Milroy into three distinctive lithologies. The upper and lower Milroy characterized by "ribbon" dolostones and limestones are separated by an argillaceous zone of stromatolites, ooids, and a restricted fauna. The lower and middle units were interpreted, above, as the HST of sequence M1A, while the upper Milroy is here considered to be the TST facies of the M1B sequence. It is nearly identical to the lower Milroy in facies appearance; however, the upper Milroy shows an increased abundance of dolomitic limestone over coarser-grained dolostones, and many of the ribbon limestones are slightly thicker owing to increased bioturbation that appears to destroy some of the thinner ribbons. Meter-scale cycles in this unit are difficult to identify and most have stylolitic contacts separating them. In general, they appear almost aggradational in nature, except that bioturbation becomes slightly more significant upward with some cycles showing increased development of firm- to hardgrounds. It is clear that this facies is extremely restricted, even more so than underlying Milroy. This suggests that the Adirondack Arch may have become a much more pronounced feature that isolated, at least partially, the central Pennsylvania area from more open marine circulation. Without many other depositional signatures, the TST is interpreted here by the lack of siliciclastics, and the pattern of increased bioturbation.

The HST of the M1B sequence is recognized within the Clover Member of the Loysburg. The base of the Clover was placed below the lowest bed with abundant and diverse fauna and above the last of the relatively thick dolostones Chavetz, (1969). It is typically a sublithographic ostracod-bearing limestone with dark-gray silty and argillaceous beds separating off more distinct meter-scale cycles. Bioturbation is clearly more abundant in this unit and the “ribbon beds” of the Milroy are no longer as distinct due to bioturbation; however, they become predominant once again toward the top of the Clover. Kay (1944) recognized occasional channeling and a couple of “intraformational conglomerates” in this unit and these appeared to be more prominent to the southeast toward the Adirondack Arch. These conglomeratic beds appear to have formed in response to possible faulting and erosion along the margin of the Adirondack Arch at this time. Unlike elsewhere, the M1B HST is not characterized by a dominant shale facies, although siliciclastic event 2 is likely represented by the silty and argillaceous partings found in the upper Clover. The upper SB of the M1B sequence in PA is likely recorded in the interval of the intraformational conglomerates, but as there appear to be multiple levels of conglomerates, it is difficult to place a specific contact to represent the SB. The immediately overlying Eyer Member of the Hatter Formation is distinct in that it is composed of cycles of medium- to thick-bedded skeletal grainstones and micrites with only minor stylolitic to shaly partings, these appear to be developed as thin broad lenses, deposited in low-lying areas. The Eyer, a relatively thin-unit, and the overlying Grazier Member (the middle member of the Hatter Formation) onlap the Adirondack Arch to the southeast. This interval suggests initiation of the M2 TST and helps constrain the underlying Milroy to the previous sequence.

M2 Sequence: Carntown-L. Pamelia-Hatter-L. Shippensburg

The M2 sequence (**figure 10**), first described by Holland and Patzkowsky (1996, 1998),

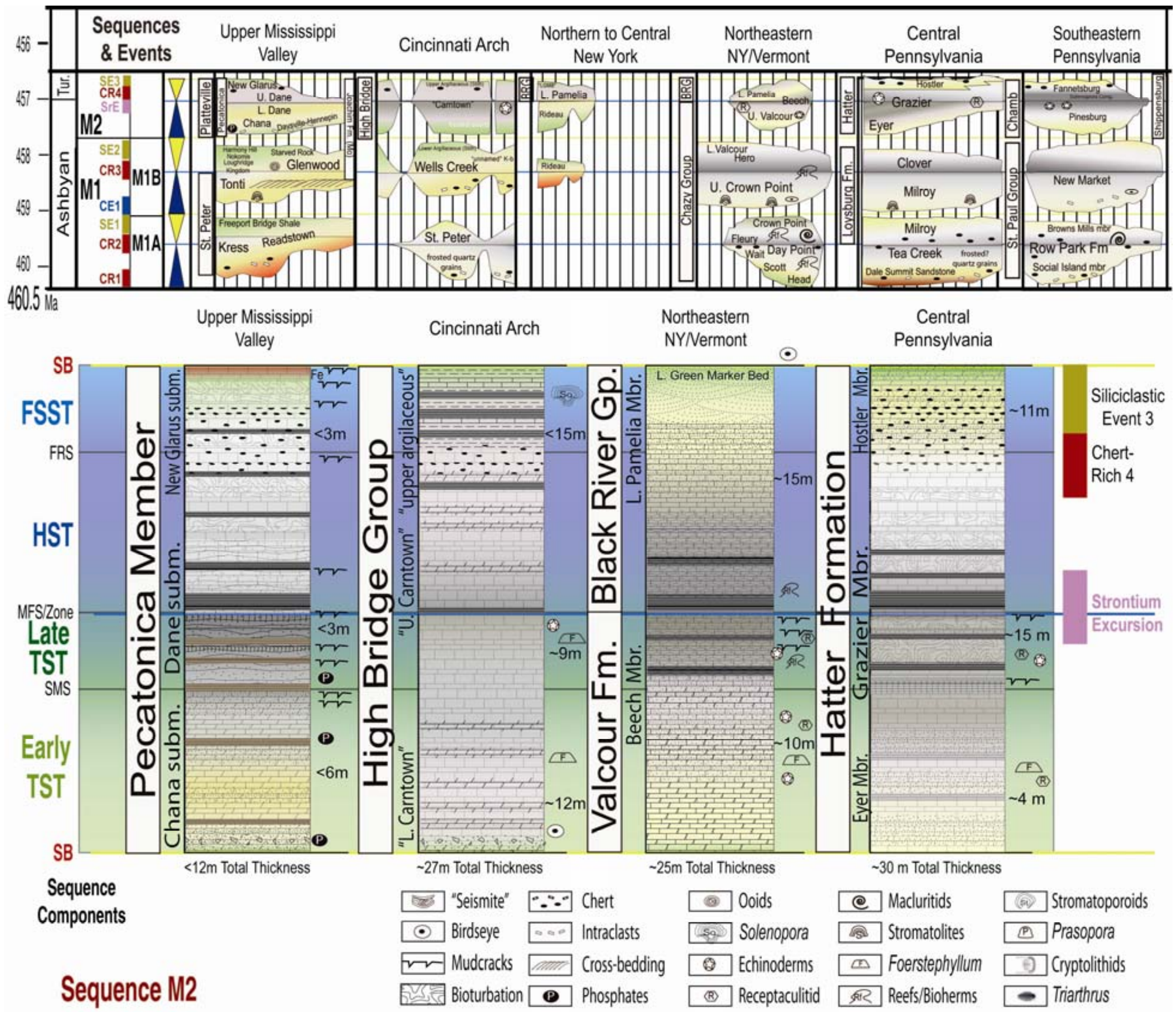


Figure 10: Correlated framework of the M2 depositional sequence (Ashbyan-Turinian) for the Upper Mississippi Valley, Cincinnati Arch, Northeastern New York, and central Pennsylvania. This sequence straddles the traditional Chazy-Black River Group boundary.

was recognized as the first sequence of the "Black Riveran" and was defined by the Pierce and Ridley Formations of the Nashville Dome of Tennessee. This sequence was correlated with the upper Camp Nelson Formation of the High Bridge Group and the overlying Oregon (**see figure 6**). As documented in **figure 7** and elsewhere, these correlations have been modified here. Moreover, in New York State, the first depositional sequence of the Turinian ("Black Riveran") is interpreted here to straddle the Chazy-Black River Group boundary although the stratigraphic and nomenclatural differences are complex.

SB-TST

In the Champlain Valley region the lithostratigraphic boundary has not been well established between the Black River and the Chazy, and has been viewed as transitional (Fisher, 1968, see chapter 3). In contrast, to the northwest in Quebec, the boundary was described as the abrupt change from the Laval Formation dolostones (Chazy-equivalent) into the basal Pamelia sandstones and dolomitic limestones, and is seen as unconformable (Salad Hersi & Lavoie, 2001). In this case, Salad Hersi and Lavoie interpret the Pamelia of the Quebec region to be equivalent of only the uppermost Pamelia and therefore the unconformity in that region is likely a composited sequence boundary with little to no evidence of the M2 sequence in the Beauharnois Arch region – it was apparently uplifted at this time and physically isolated the Ottawa Embayment from the St. Lawrence Platform.

In the type Chazy area southeast of the Beauharnois Arch, where deposition was more continuous, the M2 sequence is more fully developed. Given the development of the Beauharnois Arch at this time, facies in the M2 sequence again show significant lateral change with shallow-water, siliciclastic dominated facies in the type Chazy area, and significantly more carbonate in the eastern Champlain area of Vermont. Nonetheless, the M2 sequence TST is represented by the uppermost Valcour Formation (Chazy Group) and the M2 sequence HST by the lowermost Pamelia Formation (Black River Group). In some cases, the uppermost Valcour as defined may actually be synonymous with the lowermost Pamelia Formation. The Valcour-Pamelia contact as drawn by Fisher (1968) in northeastern New York is tentatively considered herein to be a MFS rather than an unconformity. The SB of the M2 sequence is drawn in the base of the Beech Member of the Valcour Formation, which appears to onlap the underlying M1B sequence, although it is not as laterally extensive.

In its type area (Grand Isle, Vermont) the lower Beech Member is about ten meters of somewhat dolomitized grainstones and crinoid and brachiopod encrinites that become less dolomitic upward. These grade-upward into shaly nodular silty wackestones and packstones containing rather large colonies of bryozoans developed on several hardground surfaces. Upward, these beds containing large colonies grade into beds containing smaller, but relatively intact thickets of ramose bryozoans and small, well-preserved brachiopods. The top of the Beech Member is not well-exposed in the type section, but on Valcour Island to the west of Grand Isle, it is described as returning to medium- to coarse-grained, quartz-rich, cross-bedded grainstones containing numerous algal remains and receptaculitids. These in turn, are succeeded by fine-grained, birdseye micrites of typical Black River lithology that likely represent the base of the next sequence.

In New York, closer to the Adirondack-Beauharnois Arch, the Beech Member is more siliciclastic-rich and is composed of fossiliferous, medium-grained, argillaceous dolostones with lesser amounts of greenish-gray quartz siltstones and minor reddish sandstones. Although these lithologies are typical of the Pamelia, these beds are still recognized as Valcour in that they contain many of the fossils found in the type Valcour. Moreover, the uppermost beds of the Valcour show a transition from dolomitic limestones into peritidal birdseye micrites similar to those found above the Valcour in the eastern Lake Champlain region.

Overall, it appears that the Beech Member, in the eastern Lake Champlain area, may itself represent the entire M2 depositional sequence with the coarse-grained dolomitic interval representing the TST, and the bryozoan-bearing hardground interval representing the MFS. In New York, west of Lake Champlain, the details of the sequence are less clear and in need of more study, but the lower Beech TST is significantly more dolomitic and argillaceous than further east and onlaps the Hero Member of the Valcour Formation. This unit is correlated across the Adirondack/Beauharnois Arch with the upper Rideau Sandstones

MFS-HST

The silty argillaceous bryozoan-bearing wackestones and packstones of the Beech Member represent the early HST of the M2 sequence in western Vermont. The late HST or RST interval is recorded by the shallow, shoaling facies (cross-bedded grainstones) below the first birdseye micrites. In eastern New York, the HST contains more siliciclastic sediment (siliciclastic event 3) indicative of its more proximal location and indicative of lower Pamelia lithologies that appear to prograde eastward over the carbonate-dominated TST. In northwestern New York, and Ontario, this interval is referred to as the “lower green marker bed” that marks the top of the lower Pamelia Formation. The transition to the first relatively pure-carbonates (peritidal micrites) represents a renewed period of base-level rise of the M3 sequence coincident with the middle Pamelia Formation.

CORRELATION OF THE M2 SEQUENCE

<u>Upper Mississippi Valley</u>

In the Upper Mississippi Valley, Witzke and Bunker (1996) recognized the lowest Transgressive-Regressive cycle of the “Platteville sub-cycle” as cycle 2. They identified the unconformity bounded Pecatonica Member (Formation in Illinois) of the Platteville as representative of the first “Black Riveran” cycle. Here the Pecatonica is considered the equivalent of the M2 sequence. The Pecatonica has been subdivided into several units especially in the Illinois to Missouri region. Overall it comprises a dark, interbedded dolostones and fine- to coarse-grained limestones separated by thin, brown, organic-rich shales (Nelson 1996). The lower Pecatonica Member (Chana sub-member) is between 2 and 6 meters thick and is dominantly a brown, thick-bedded, vuggy, pure dolostone that contains disseminated quartz grains and occasional ribbons of well-rounded and abraded quartz sands (Willman et al., 1975). The very base of the Chana is occasionally described as argillaceous and similar to the underlying Glenwood Formation, except that it often carries small clasts and pebbles including some that appear to be phosphatic. Templeton and Willman (1952) had previously recognized a thin unit referred to as the Hennepin Member of the Pecatonica, but it appears that this unit is a thinner lateral equivalent of the Chana. The middle to upper Pecatonica is represented by much more fossiliferous dark gray occasionally laminated limestones and interbedded brown laminated dolostones. This interval (referred to as the Dane sub-member) contains brachiopods, ostracods, and a variety of algae, especially in some of the meter-scale cycles showing slightly thicker dark-gray shales (upper Dane sub-member). These beds become more bioturbated and the upper Pecatonica, the New Glarus sub-member, contains nodular chert horizons (chert-rich interval 4 of this study).

In terms of sequence stratigraphy, the Pecatonica appears to onlap the Wisconsin Arch from the south and west from the Illinois Basin. Witzke and Bunker (1996) suggested the phosphatic zone and brown, organic-rich shale (Chana) facies of the Pecatonica indicate evidence for condensation during marine transgression. The dolostone-bearing phosphatic pebble lag likely represents a combined SB through TST interval, and the incorporation of quartz-sands in the Chana, likely reflects the last stages of wind-blown quartz sands during the M2 sequence TST. These former authors also recognized two prominent, closely-spaced, regionally extensive contacts, at the top of the Pecatonica (New Glarus sub-member). Interpreted as hardgrounds or transgressive discontinuities or solutional disconformities by Witzke and Bunker, these “hardgrounds” likely represent the sequence boundary interval of the M2-M3 sequence. It remains to be seen if these represent a FRS and SB or a SB and transgressive surface respectively.

Cincinnati Arch

In the Cincinnati Arch region, the TST of the M2 sequence is essentially recognized as the “Carntown” unit of the Camp Nelson recognized by Stith (1979). The Carntown is located above the “Lower Argillaceous” unit (siliciclastic event 2 at the end of the M1B sequence – top Wells Creek) and below the “Upper Argillaceous” unit (siliciclastic event 3 at the end of the M2 sequence). The uppermost Carntown is correlative of the lowest High Bridge Group exposed in the central Kentucky region, where the Upper Argillaceous unit is correlated with the “open marine shale marker” of Ettensohn (1992). The “open marine shale” was evidently a slightly deeper facies, at least in the Kentucky River fault zone, than most time-equivalent facies, including those seen northward into the subsurface of Ohio where more typical poorly fossiliferous dolomitic facies are developed. Local movement on faults in the region may explain the slightly deeper facies in this region. As discussed in chapter 5, Ettensohn (1992), suggested that this interval represented a period of flexural relaxation in the development of the Blountian tectophase.

Noger and Drahovzal (2005) correlated the high-calcium Carntown interval from the eastern side of the Cincinnati Arch with the Pecatonica of the Illinois Basin. They favored use of the term Pecatonica in the subsurface west of the Cincinnati Arch as a new member of the Camp Nelson Formation, but herein the interval is referred to as the Carntown unit. In the Cincinnati Arch region, much of the Camp Nelson is composed of numerous stacked, peritidal, meter-scale cycles dominated by pure micritic “birdseye” limestones. The lower Carntown is composed of very pure peloidal micrites with minor dolomitic zones and the upper Carntown is a ribbon facies of interbedded dolostones and dolomitic limestones immediately below the upper argillaceous interval. As discussed in chapter 5, Stith (1979) interprets the Carntown as a shallowing-upward succession overall. Very few fossils have been described from the Carntown except for some corals including *Tetradium* and a larger tabulate – similar to *Foerstephyllum* have been noted in cores within the lower Camp Nelson. In addition to corals, the “open marine shale” of Ettensohn includes a fauna that includes brachiopods, bryozoans, ostracods, and gastropods which may reflect a somewhat restricted fauna.

Although few, details of this depositional sequence suggest that the coral-bearing lower Carntown represents the TST that onlaps the erosion surface and the underlying Wells Creek. The HST is represented by the inter-bedded ribbon limestone facies that Stith interpreted as a supratidal succession. The HST facies is also shown to have a greater insoluble quartz-silt fraction and near the top of the succession abundant algal nodules are recognized below the level of the uppermost shale. Given the observation that the Carntown pinches out against topographic highs in some limited areas (Stith, 1979) and is succeeded above by what has been interpreted as the late HST siliciclastic event 3, there is at least some evidence that it represents a depositional sequence similar in scale and magnitude to the Pecatonica. The SB at the top of the

M2 sequence is established above the upper argillaceous marker and below the first intraclastic beds of the M3 sequence.

Central Pennsylvania

In the Ridge and Valley, the M2 sequence is here correlated with the Hatter Formation. The Hatter is composed of three members (Eyer, Grazier, and Hostler). The lowest Eyer Member is regionally restricted in its occurrence and rests sharply above the underlying Loysburg (Clover Member). The contact between the two is considered the SB of the M2 sequence. The Eyer is typically a fossiliferous, coral-bearing, medium- to thick-bedded grainstone interval interbedded with medium-bedded calcilutites with thin, irregular stylolitic to shaly partings separating meter-scale cycles. Some of the finer-grained grainstones and calcilutites are dolomitized, but the unit contains a fauna dominated by thin branching bryozoans, ostracods, and a few small brachiopods. The Hatter Formation becomes more argillaceous and finer grained upward in the overlying Grazier Member and this succession was interpreted as an upward-deepening or transgressive unit by Kay (1944).

The Grazier is dominated by massive-bedded, bioturbated mudstones and wackestones separated by shaly nodular partings, the non-carbonate fraction increases to nearly 20% in the upper Grazier. Occasional thin grainstone stringers are observed in the lower Grazier and one bed is preserved as the shallowing-upward cap of a meter-scale cycle. In this case, the rippled bed has vertical burrows filled with chert and may represent a hardground surface near the middle of the TST. A second dark-stained contact sits above the bioturbated wackestones of the lower Grazier, and below the darker, wavy- to planar-interbedded calcisiltites and wackestones of the upper Grazier. This second contact is likely another hardground and maybe coincident with the MFS, and the increase in siliciclastic detritus in the upper Grazier may record the subsequent HST facies of the cycle. Overall, the Grazier Member has a slightly more diverse faunal assemblage including bryozoans, large cephalopods and receptaculitid algae.

The topmost member of the Hatter Formation is the Hostler Member and is known for its highly siliceous composition. The wavy-bedded wackestones and calcisiltites of the Grazier Member grade upward into dolomitic laminated calcisiltites which in turn are sharply overlain by fossiliferous, coarse-grained grainstones, packstones, and bioturbated wackestones suggesting a shallowing over the middle Hatter Formation. These beds are the most fossiliferous of the Hatter and are highly diverse with up to 28 different taxa reported (see chapter 4). This middle Hostler Member is finally capped by a greenish, planar-bedded dolomitic limestone unit that contains upward of 37% insoluble residues composed of quartz silt and disseminated chert. Here this facies is considered representative of the late HST/RST facies of the sequence and is correlated with chert-rich interval 4 and siliciclastic event 3 elsewhere. The next overlying sequence is floored by major conglomerates and intraformational breccias indicating a new sequence boundary interval immediately above the Hostler.

M3 Sequence: U. Camp Nelson–Pamelia–Benner-Doylesburg/Housum

The M3 sequence, which in Tennessee is represented by the Lebanon Limestone, is here correlated with the middle Pamelia Formation of New York and Ontario. It is the first, third-order depositional sequence completely in the Black River Group of the type region. The M3 sequence is well-known for its echinoderm faunas and has been studied in some detail in the southern Appalachians. It was previously correlated with the lower Tyrone Formation of the Cincinnati Arch. In New York, the M3 sequence records the first major marine flooding event and incursion of carbonates into the type Black River region of northwestern New York State (**figure 11**).

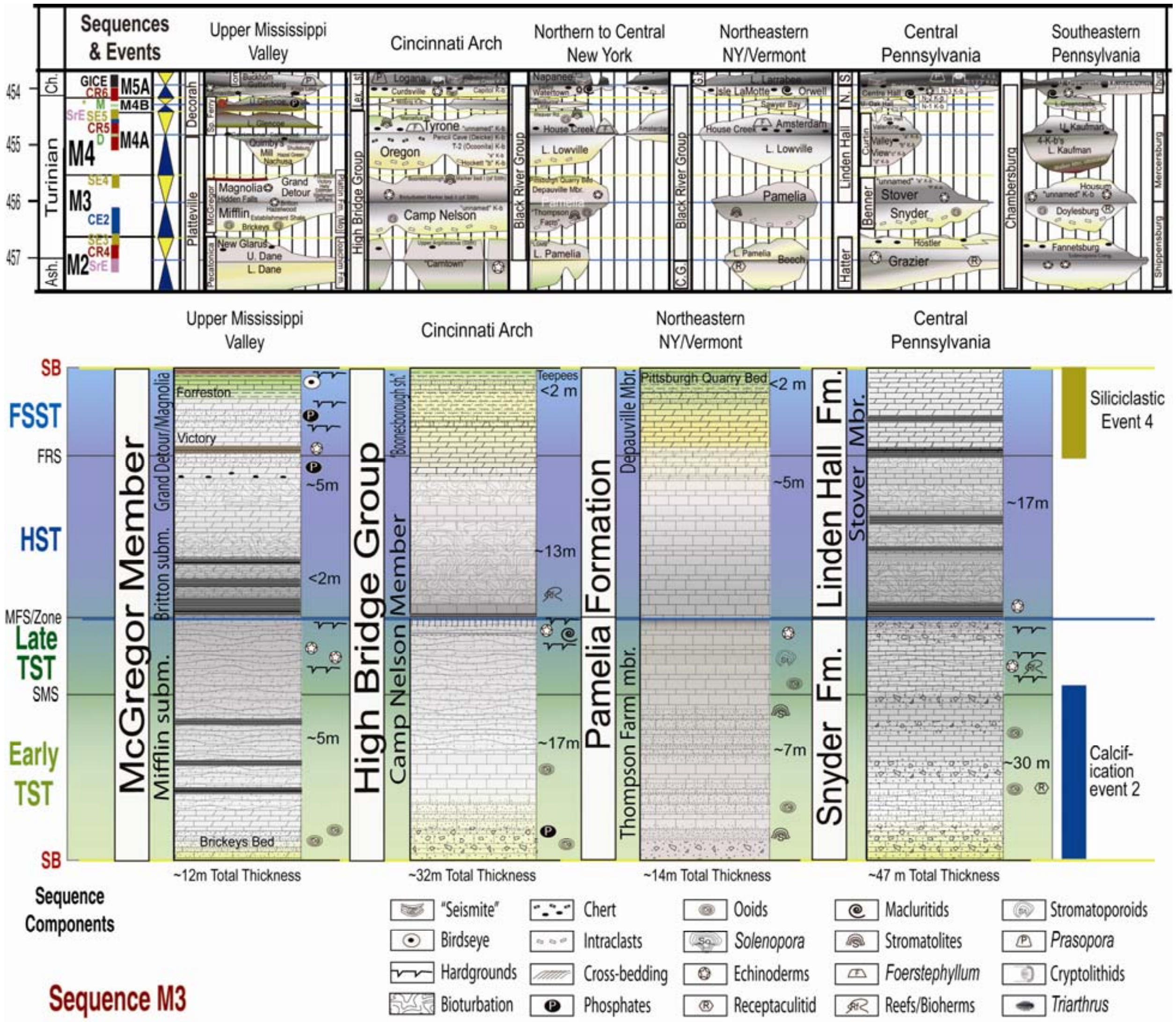


Figure 11: Correlated framework of the M2 depositional sequence (Turinian) for the Upper Mississippi Valley, Cincinnati Arch, Northeastern New York, and central Pennsylvania.

SB-TST

In the type region of the Pamelia Formation, the basal SB of the M3 sequence is recognized between the top of the lower Pamelia and the base of the middle Pamelia. The latter member (referred to herein as the "Thompson Farm" member), onlaps and erosionally truncates portions of the lower Pamelia. The SB is easily recognized where the middle Pamelia onlaps the

Precambrian basement and includes occasional reworked fragments of Grenville-aged basement rock and minor feldspathic sandstones derived from that unit. Elsewhere, the SB interval contains thin quartz-rich siltstone beds and argillaceous desiccation-cracked micrites that are immediately superseded by moderately fossiliferous coarse-grained, micrite-rich wackestones representative of the TST. In some places, the Thompson Farm shows minor evidence of channeling and at its base contains intraclasts including shale rip-up clasts and dolostone fragments that appear to be derived from the underlying lower Pamelia and/or from the Theresa Dolostone.

Especially in the southern Black River Valley, the Thompson Farm member is typically found to contain crinkle laminated facies that grade into LLH to domal stromatolites and ooid-rich mudstones to wackestones (oomicrites) often bearing *Tetradium* corals. Northwest of the Adirondack Arch, toward the Kingston Trough, the ooid and stromatolite facies of the TST are less well developed and are replaced laterally by fossiliferous micrites (wackestones and occasional packstones) containing bryozoans, cephalopods, brachiopods, stromatoporoids, crinoids, and a variety of bivalves. Meter-scale cycles in this interval were recognized and correlated by Cornell (2001). This facies also contains abundant peloids ranging in size from a few millimeters up to a centimeter in diameter. Many of the larger ones appear to be derived as clasts of early, but incompletely cemented micritic sediments that were remobilized and redistributed by storm processes. Other peloids of this unit may be algal in origin, especially in the case where some show a central nodule or skeletal fragment surrounded by a darker rim of structureless micrite. These facies are uncharacteristic for much of the Black River Group and this occurrence is herein referred to as representative of calcification event number 2. Development of these facies may have coincided with the later stages of the TST when the region was influenced by the influx of more normal marine water and storm activity helped mix somewhat restricted water masses with differing carbonate saturation states.

MFS-HST

The maximum flooding interval of the M3 sequence in New York State is developed within the upper Thompson Farm member and below the base of the Depauville Member. Its details have not been explicitly established due to poor exposures in the type area. Where the contact is exposed in the Thompson Farm Quarry at Lafargeville, it occurs above the quarry floor and is not easily accessible for detailed study. Nonetheless, the upper Thompson Farm is slightly more bioturbated than the lower half of the formation and appears to be arranged into several meter-scale cycles. It transitions upward gradually into less-fossiliferous, more siliciclastic-rich, dolomitic facies, which is the Depauville Member of the Pamelia Formation. The Depauville shows evidence for shallowing-upward cycles, each becoming progressively shallower and showing evidence for pronounced supra-tidal facies and more substantial dolomitization below the level of the sandy shales and siltstones of the Pittsburgh Quarry Bed that caps the sequence. The upper Depauville shows evidence for early primary dolomites, along with evaporitic textures including evaporitic pseudomorphs and celestite-filled vugs, indicating a period of extreme restriction and evaporation that followed more-open marine conditions. Thus this succession is interpreted as the HST interval through the base of the Pittsburg Quarry Bed.

In contrast to Cornell (2001) wherein the sequence boundary was placed at the base of the Pittsburgh Quarry Bed, it is here suggested the upper SB of the M3 sequence is actually at the top of the Pittsburgh Quarry Bed. The sharp, irregular (and channelized?), pyrite-stained contact at the base of the Pittsburgh Quarry Bed (SB sensu Cornell, 2001) is re-interpreted as a possible FRS surface. The Pittsburg Quarry Bed, itself, is a greenish, laminated siltstone, and mud-cracked dolomitic limestone to dolostone unit with small, well-rounded, polycrystalline quartz grains (siliciclastic event 4) that is interpreted here as the latest HST/RST deposit sitting below the regional sequence boundary surface and the base of the overlying Lowville Formation.

CORRELATION OF THE M3 SEQUENCE

Upper Mississippi Valley

In the upper Mississippi Valley region, the M3 sequence corresponds with the McGregor Limestone Member of the Platteville Formation. Correlation of this sequence with the M3 sequence of Tennessee is supported by the echinoderm faunas studied by Kolata (1975) from the Grand Detour that correlate well with those recognized in the Lebanon Limestone. The Platteville Formation has been subdivided into a substantial number of units across Iowa, Illinois, Wisconsin and Minnesota. Witzke and Bunker (1996) previously recognized the McGregor Member as their third, third-order transgressive-regressive cycle (TR-3) of the St. Peter-Platteville interval. TR-3 was internally subdivided into two smaller cycles, A (lower), and B (upper). Ludvigson and colleagues further formalized these depositional sequences and identified three sequences (the Mifflin, Grand Detour, and Nachusa Sequences). These are partially reconstructed here to form two, third-order depositional sequences as discussed below.

The lowest unit of the McGregor Member is the Mifflin sub-member and is described as a dolomitic wavy- to nodular-bedded limestone often showing thin shale partings between some bedding planes. Overall the Mifflin is a micrite-dominated wackestone characteristic of low-energy inner shelf environments. In Illinois and Missouri, the lower sub-unit of the Mifflin, the Brickeys Limestone, is an oolitic limestone facies and is here recognized as representative of calcification event 2. Both facies of the Mifflin grade upward into more fossiliferous packstone lithologies and even echinoderm-rich grainstones suggesting possible shallowing (top of 3A sub-cycle as per Witzke & Bunker), or as suggested here, shoaling conditions as water depths increased into the later TST allowing for increased wave development and winnowing. A prominent surface below the uppermost Mifflin (Briton sub-member of Kolata, 1975) is interpreted as a minor discontinuity that shows evidence of hardground development and minor sediment starvation. In Iowa and Minnesota this contact sits below the level of the shaly Hidden Falls Member. Eastward into northern Illinois and Wisconsin, the equivalent contact is recognized at the top of the Hazelwood Limestone sub-member and at the base of the Briton sub-member of the Mifflin Member (Kolata, 1975). In this region, the unit is less shaly and more wavy-bedded and contains stringers of wackestones, packstones, and fine-grained grainstones containing well-preserved echinoderms and crinoid holdfasts attached to some hardground bedding planes. The contact at the base of the Briton sub-member is likely developed in the latest TST during a period of rapid sea-level rise, and the true MFS of sequence M3 may be developed slightly higher at the level of one of the holdfast-bearing, grainstone beds.

The Hidden Falls Member / upper Briton sub-member, and the overlying lower Grand Detour Members (Magnolia equivalents) in turn, represent the early HST of the M3 sequence and the base of the second sub-cycle of the McGregor as defined by Witzke and Bunker (1996). Ludvigson and colleagues (2004) suggested this interval represented its own depositional sequence with a basal condensed interval and an overlying fossiliferous carbonate succession. Here this interpretation is slightly modified. The Hidden Falls is typically an argillaceous, dolomitic limestone and bentonitic shale unit and is likely an early HST facies. As suggested by Witzke and Bunker (1996) this unit is likely representative of a progradational facies developed during shallowing of the TR-3 cycle and regional base-level drop. Incidentally, Sloan (1956) suggested the unit may have been deposited concurrently with local basement tectonic activity at this time in the region of the Wisconsin Arch that may have contributed additional sediment. In the Platteville type region, the upper McGregor is composed of the Magnolia Member which is similar lithologically to the Hidden Falls Member except that it has fewer interbedded shales owing to increased bioturbation of the unit. The Magnolia and its equivalent the Grand Detour (of the Illinois Basin) are complexly developed and contain a number of local phosphatic lags and brown, organic-rich shales, occasional chert horizons, and echinoderm-rich grainstone beds. Witzke and Bunker indicate that the interval shows evidence of condensation and higher-order cyclicity (meter-scale cycles?), but overall suggest a general shallowing-upward sedimentary sequence capped by coralline algal facies and even peritidal mud-cracked facies. Here the TR-3 cycle B of Witzke and Bunker is considered the HST of the M3 sequence.

In Illinois, the upper Grand Detour is represented by the Victory and Forreston sub-members the latter of which is abundantly argillaceous, green dolomitic limestones, and buff-weathering dolostones, and the former is typically a peritidal white calcilutite with algal laminations and fenestral micrites. Locally, in Iowa, as suggested by Ludvigson and colleagues (2004), the Victory Member is corroded (karstified?) and truncated below the Nachusa Formation and the base of the M4 sequence above. The return appearance of the argillaceous interval at the top of the Grand Detour/McGregor Member constitutes the equivalent of siliciclastic event 4 as defined elsewhere and represents the very latest HST/RST phase of the M3 sequence. As with the Pittsburgh Quarry Bed in New York and Ontario, the sharp contact at the base of the Victory sub-member is interpreted as a possible FRS. The sharp upper contact between the Forreston sub-member and the overlying Nachusa is thus considered the SB surface, although the SB appears to truncate the Forreston and portions of the Victory sub-members in some areas.

Cincinnati Arch

In the Cincinnati Arch region, the M3 sequence is composed of the upper Camp Nelson Formation. This portion of the formation is constrained between the top of the upper argillaceous interval (of Stith, 1979) and the top of Boonesborough Shale or Marker Bed I (of Stith). This portion of the Camp Nelson thus contains the extensively bioturbated "Marker Bed II" of Stith and is more fossiliferous than the underlying M2 sequence in the same region. Overall, like much of the High Bridge Group, this interval is dominated by shallow subtidal to peritidal lithographic to sub-lithographic limestones. However, some important facies changes are recognizable and permit sequence stratigraphic assessment. The SB interval of the M3 Sequence is recorded by a zone of minor intraclasts and occasional algal (?) peloids (similar to those recognized in the Pamelia Formation of New York). Some of the irregular peloids in this zone are bluish-black in color and are possibly enriched in phosphate. In the eastern Jessamine Dome region, these basal beds grade upward into slightly coarser-grained facies that are more similar to the Lebanon Limestone of Tennessee than to the exceptionally fine-grained micritic limestones of the typical High Bridge Group further to the west. In many cases, this zone is heavily stylolitic owing to the return to exceptionally pure limestones (compared to the HST of the M2 sequence below). These slightly coarser facies appear to represent the same shoaling-upward pattern recognized in the Mifflin Member of the Platteville Formation of the Upper Mississippi Valley.

The MFS-HST interval of the Camp Nelson coincides with the transition into wavy-bedded micrites and dolomitic wackestones that contain the extensively bioturbated facies recognized by Stith. Just below the sharp change into HST facies, are a number of hardground surfaces that show evidence of vertical borings, small colonies of *Tetradium*, and relatively planar contacts. The interval also contains an unnamed K-bentonite and several condensed packstone pavements showing an abundance of brachiopods, cephalopods, and large macluritid gastropods on some bedding planes. The condensed nature of these beds suggests this interval likely represents the maximum flooding zone and an interval of sediment starvation. Taphonomic signatures of some of the cephalopods suggest they may have been reworked during the latest TST.

HST facies of the M3 sequence are characteristically heavily bioturbated, massive calcilutites, and calcisiltites (micritic wackestones). Few shale partings are recognized owing to the intensity of bioturbation. However, some of the cycles show occasional large coral colonies, echinoderm fragments, as well as other fossil taxa at their bases, and then shallow upward. As in New York and the Upper Mississippi Valley, high-frequency, meter-scale cycles show a shallowing-upward pattern overall. In these HST facies, meter-scale cycles show a gradual increase in the thickness of peritidal to supratidal dolomitic limestones and dolostone cycle caps suggesting the shallowing. As elsewhere, the top of the M3 sequence (latest HST/RST) is marked by the sudden appearance of a siliciclastic-rich interval (SE-4) as recognized here as the green Boonesborough Shale Marker. This unit was previously interpreted to contain “teepee” structures which may indicate the development of extensively restricted and possible evaporitic facies at this time immediately prior to deposition of the Oregon Formation that records the onset of the M4 sequence.

Central Pennsylvania

In Central Pennsylvania, the M3 sequence is essentially equivalent to what was referred to by Kay (1944) as the Benner Limestone Formation. As discussed in chapter 4, the Benner Formation is not consistently recognized as a lithologic unit in Pennsylvania. Nonetheless, the Benner Formation contains the Snyder and Stover Members. Some workers today consider the Snyder (here interpreted as the TST of the M3 sequence) as its own formation, and the Stover (HST of the M3 sequence) is generally included as the basal member of the Linden Hall Formation.

The Snyder Member (Formation) is a white- to light-gray-weathering limestone characterized by its prominent limestone pebble conglomerates, well-developed ooid packstone to grainstone beds and abundant fossil fragments and algal (*Solenopora*/*Girvanella*) rudstones. The development of the ooid-rich interval is characteristic of this sequence and recognized as representative of calcification event 2. The Snyder onlaps older, underlying units to the southeast of the type region on the Adirondack Arch and represents a deepening facies overall. The top of the Snyder (latest TST to MFS interval) is capped by the widespread and most significant intraclastic zone recognized. These beds are typically associated with three to four hardground surfaces within a fossiliferous, moderately coarse-grained interval. The presence of these intraclastic beds was interpreted by Laughrey and others (2004) as representative of a late HST to sequence boundary interval; however, it is more likely that these intraclasts are related to storm-influenced erosion and transport of previously cemented hardgrounds. Moreover, the large number of bioherms and biostromes in this interval suggest these are deeper facies than assumed. However, in some cases, fragments of limestones and dolostones similar to the underlying Hatter Formation are observed in this facies. The occurrence of these “extraformational” intraclasts can be explained by the pattern of onlap of the Snyder Formation onto the Adirondack Arch to the southeast and erosional transport of those lithoclasts during storm events during the transgressive phase. In addition, as discussed elsewhere, it is apparent that coincident with deposition of the M3 sequence, there may have been tectonic modification over many intracratonic features such that the Adirondack Arch may have uplifted slightly at this time allowing underlying materials to be reworked and deposited in the central Pennsylvania graben to the northwest.

The HST of the M3 sequence in Pennsylvania is recognized as the Stover Member of the Linden Hall Formation. The sequence is characterized as an extensively bioturbated, dark gray, massive-bedded, fine-grained calcilutite to wackestone facies and is similar to what is recognized in the same sequence in the Cincinnati Arch region and in the Upper Mississippi Valley region. It also shows evidence upward for increased wavy lamination and partially dolomitized limestone facies interbedded with shaly seams through the top of the unit. Some of the clay-rich seams are considered to be K-bentonites. This succession thus suggests a shallowing-upward pattern and therefore the M3 HST. Kay (1944) showed that the Stover itself was substantially thinned, and eventually truncated below superjacent units along the Adirondack Arch. This pattern suggests that the upper contact of the Stover Member with the Valley View Member of the Linden Hall Formation (Curtin Formation of Kay, 1944) is the M3-M4 SB. In the Laughrey and colleagues (2004) sequence scheme, the Stover is not recognized, but is included in their Linden Hall. The Stover is likely equivalent to what they referred to as “mining unit 17” and interpreted as a TST to early HST facies. However, the bioturbated wackestones and ribbon grainstone beds do not appear to be condensed or reworked, but suggest evidence for rapid deposition including normally-graded storm beds capped by occasional rippled surfaces interpreted here as a shallowing succession and thus an HST interval. These eventually are capped by another intraclastic conglomerate that marks base of the next sequence, and another zone of intraclastic limestones interbedded with coarser-grained, hardground-bearing packstone and grainstone dominated interval.

The Stover itself is a characterized faunally as a sponge-echinoderm-bearing unit which allows for its correlation to the Lebanon Limestone of Tennessee, the McGregor Limestone of the Upper Mississippi Valley, and the middle Pamelia Formation of New York. Evidence for siliciclastic event -4 at the end of this cycle is not present in this western Ridge and Valley region and it appears that the region was completely isolated from this siliciclastic influence (except for occasional K-bentonite deposits). However, to the southeast across the Adirondack Arch into the Cumberland Valley, the equivalent interval does contain a quartz-rich siltstone interval that may represent the end of the siliciclastic event and the initiation of the next sequence.

M4 Sequence: Oregon/Tyrone-Lowville-Curtin-U. Mercersburg

The M4 sequence of the Nashville Dome was represented by the Carters Limestone which was widely recognized to contain both the Deicke and Millbrig K-bentonites. This succession was, in turn, correlated with the Tyrone Formation of the High Bridge Group of central Kentucky (Holland & Patzkowsky, 1996, 1998). Here it is suggested that the M4 sequence (**figure 12**) is represented by the Oregon (TST) to Tyrone Formation (late TST to HST) of the latter region. It is also correlated here with the Lowville Formation of New York, and the upper Linden Hall Formation (Curtin Formation) of central Pennsylvania. Here the uppermost part of the M4 sequence is distinguished and recognized as a “mini-sequence” (termed the M4B sequence see below). Although it likely represents the last parasequence of the M4 Late

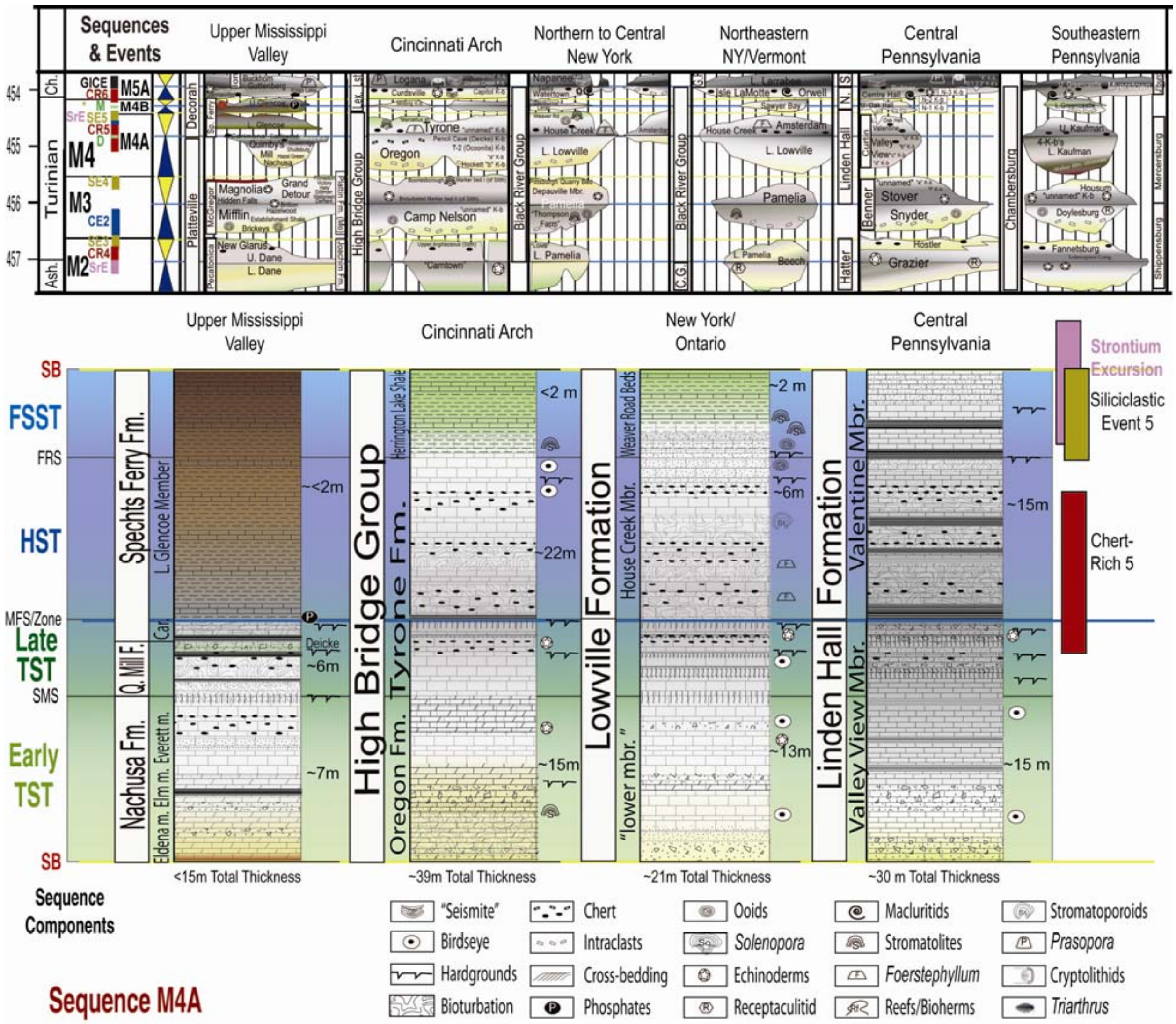


Figure 12: Correlated framework of the M4A depositional sequence (Turinian) for the Upper Mississippi Valley, Cincinnati Arch, Northeastern New York, and central Pennsylvania.

HST/RST, its behavior and lithologic range is more typical of a depositional sequence and is atypical compared to subjacent and superjacent small-scale cycles. As the M4B contains the Millbrig K-bentonite, it is distinguished here as it may reflect an as yet, unrecognized but important, short-term sea-level rise and fall event leading into the M4-M5 SB.

SB-TST

The sequence boundary of the M4 sequence in the type Black River region of New York is represented by the prominent lithologic break between the Pamelia Formation and the overlying Lowville Formation. The contact and sharp change from the siliciclastic-rich, Pittsburgh Quarry Bed into the relatively pure carbonates and dolomitic limestones of the Lowville Formation demarcates a new period of base-level rise. Siliciclastic sediments, characteristic of the latest M3 sequence, are once again sequestered and source areas are inundated by marine transgression. Moreover, it appears that much of the Adirondack Arch, the Beauharnois Arch, and much of the southern Canadian Arch region area were eventually inundated during deposition of this sequence, thus substantially reducing the supply of siliciclastics to the region. The only available siliciclastic sources were apparently located far to the north, to the west near the Transcontinental Arch, or possibly to the southwest in the Ozark Dome region. Moreover and most importantly, siliciclastics were also emanating from the south out of the developing Blountian Highlands.

The lower Lowville member (upper Gull River Formation of Ontario) shows a progressively upward deepening pattern of meter-scale cycles (**see figure 5**). Beginning above the Pittsburgh Quarry Bed, Cornell (2001) defined six 2-2.5 meter-thick cycles that were dominated by peritidal to supratidal dolomitic limestones and fenestral micrites in thin to medium-bedded successions. These contain scattered intraclastic beds, minor current- and interference-rippled facies, abundant *Phytopsis* burrows, and occasional ostracod-rich beds. These in turn grade upward into more fossiliferous (*Tetradium*-bearing) strata dominated by shallow sub-tidal to intertidal facies that lack the supratidal dolomitic facies of the lowermost cycles. This overall upward-deepening succession is interpreted as the TST interval of the M4 sequence as was suggested by Cornell (2001).

MFS-HST

The MFS of the M4 sequence is the most prominent facies dislocation surface of the entire Black River Group. The M4 MFS is developed between the lower member of the Lowville and the House Creek Member of the Lowville Formation. In Ontario this contact is synonymous with the upper Gull River Formation – base Moore Hill Member boundary. It is a readily recognized surface located approximately 60 centimeters above the MH K-bentonite of Liberty (1969). This surface juxtaposes repetitive cycles of dark brownish-gray, coral and stromatoporoid-bearing, burrow mottled wackestones of the House Creek Member over the pale gray-weathering, *Tetradium*-bearing, fenestral micrites of the lower Lowville Formation.

The early HST House Creek-Moore Hill succession was shown to have from eight to ten 0.5 to 1 meter-thick cycles that shallow upward from sub-tidal facies through occasional oolitic facies and finally into intertidal argillaceous facies at the very top of the M4B sequence. These uppermost argillaceous beds were termed the Weaver Road Beds (Cornell, 2001) and are known to contain at least two cycles with large domal stromatolites and thrombolites. This interval represents the third calcification event of the Chazy-Black River interval, although this event is typically less widespread than the previous two events and coincides not with transgressive facies, but with late HST facies. In many cases, including in the type area, the stromatolite-bearing interval shows development of channels and spur-and groove structures. These channel structures are interpreted here to indicate that these facies may have developed during the very latest HST/RST interval. These were subsequently infilled with greenish argillaceous, mud-cracked micrites, and yellow-weathering, organic-rich, papery shales that were deposited over the carbonates prior to the next base-level rise event corresponding to the M4B “mini sequence.”

Regionally, when correlated across New York and Ontario, portions of the upper House Creek and overlying Weaver Road beds become truncated below the next overlying sequence (M4B). Across much of the eastern Mohawk Valley, much of the upper Lowville Formation is truncated in the vicinity of small-scale topographic highs yielding a composited succession of shallow-water "Lowville" facies overlain by Trenton Limestones. However, in the central Mohawk Valley, well-developed lower Lowville TST and the lower portion of the House Creek HST are present suggesting that uplift, erosional truncation and karstification occurred following deposition of the M4A sequence in this region.

CORRELATION OF THE M4A SEQUENCE

Upper Mississippi Valley

The M4A sequence, in the upper Mississippi Valley coincides with the transition from the Platteville Formation into the Decorah Formation of the Galena Group. Witzke and Bunker (1996) identified the top of their TR-3 cycle and the base of their TR-4 at the top of the McGregor Limestone as recognized by a regionally extensive hardground and iron-stained "ferruginized" contact. These authors interpreted the contact as an erosional unconformity and subsequently Ludvigson and colleagues (2004) interpreted the same contact as a sequence boundary between their "Grand Detour Sequence" and their "Nachusa Sequence." Although the sequences of Ludvigson et al., are higher-order sequences than the third-order sequences recognized here, the contact is likewise considered here as a significant third-order sequence boundary with evidence for erosional truncation of portions of the uppermost Grand Detour sub-members as discussed previously. Overlying units in the Nachusa Formation show onlap of the SB where successively higher portions of the Nachusa appear to rest on the underlying Grand

Detour sequence (M3 sequence here). Moreover, it appears that the Nachusa is absent across most of Iowa and Minnesota (Templeton and Willman, 1963) where it was likely never deposited. This region is considered to be a region of sediment starvation by Ludvigson and colleagues (2004). The Nachusa and its equivalents, appear to have been mostly constrained to the Illinois Basin of Wisconsin, Illinois, and Missouri. Thus the Nachusa Formation, collectively with the Quimby's Mill Formation, are considered representative of the M4A sequence TST described herein.

The Nachusa Formation is typically composed of three members (from bottom to top: Eldena, Elm and Everett). Collectively the succession is described in northern Illinois as a fine- to medium-grained, vuggy (and evaporitic?) dolostone that grades upward and laterally to the south into thick- to massive-bedded cycles of calcilutite facies that appear very similar to the lower Lowville Member of New York. In contrast to underlying and overlying units, it is a relatively pure carbonate unit that is only slightly argillaceous, but carries occasional cherts, at least one K-bentonite layer, occasional calcarenites, and intraclastic limestone conglomerates, especially to the south of the Wisconsin Dome (Willman et al., 1975). An interval near the top of the Everett Member contains abundant burrows "fucoids" that are apparently associated with hardgrounds just below the base of the overlying Quimby's Mill Member of the Platteville Formation. Ludvigson and colleagues (2004) suggest this diastemic contact represents a sediment starved surface, rather than a subaerial exposure surface. It is interpreted here as possibly representing a surface of maximum sea-level rise that marks the base of the condensed interval going into the late TST.

Overlying the Nachusa is the Quimby's Mill. The contact between the units is described as sharp and where the Nachusa is not present (i.e in eastern Iowa), the Quimby's Mill onlaps the

Grand Detour with pronounced unconformity. The Quimby's Mill is characteristically a brown, abundantly bioturbated, conchoidal-fracturing, micritic mudstone to wackestone interval with thin, organic-rich shale partings. To the south of the Wisconsin Dome, the interval is much thicker and some of the shaly partings are greenish-gray and weather to a whitish yellow suggestive of bentonitic shales that are abundant in this sequence elsewhere. Occasional packstone to grainstone interbeds are found associated with the Quimby's Mill, as are additional thin layers of limestone conglomerates. Many of these beds are arranged into repetitive "couplets," likely sub-meter-scale cycles (Ludvigson et al., 2004). These couplets show a gradual increase in packstone beds upward. Ludvigson and colleagues interpreted this as a shallowing of facies at the top of their "Quimby's Mill Sequence." However as suggested earlier, this coarsening of facies may reflect a shoaling of facies associated with increased water depth, increased rates of winnowing, and likely represents the later TST. Interestingly these former authors indicate that the thinning of the Quimby's Mill in east-central Iowa is due to absence of lower units by downlap, not by truncation of the upper beds. This pattern however, highly suggests a lateral expansion of facies and a pattern of onlap that is typical of a TST succession rather than a pattern associated with HST or a downlap surface as suggested by these authors.

The top of the Quimby's Mill contains a well-developed, ferruginous hardground surface that often contains phosphatic pellets and nodules (Willman and Kolata, 1978). In Illinois, Kolata and others (2001) recognized this surface as a drowning surface and according to Kolata and colleagues (1996), the Deicke K-bentonite occurs roughly at the contact between the Quimby's Mill and the overlying Spechts Ferry Member (Carimona/Castlewood sub-member). Although the prominent Deicke K-bentonite is typically considered the boundary between the

Platteville and the Decorah Formations, the Castlewood/Carimona is more similar lithologically to the underlying Quimby's Mill and continues the upward-deepening trend of alternating mudstone to packstone facies recognized by Ludvigson and colleagues (2004) in the underlying Quimby's Mill. Moreover, the top of the Castlewood also contains a zone of phosphatic nodules immediately below the first significant shales of the overlying Glencoe Member. This suggests again a period of pronounced sediment starvation, submarine corrosion, and erosion prior to a major influx of siliciclastic sediments. Thus, it is suggested here that the Castlewood/Carimona represents the uppermost portion of a well-developed, condensed interval of the M4A TST. Laterally into Iowa, both the top Quimby's Mill ferruginous discontinuity and the top Carimona/Castlewood phosphatic interval merge downward with the M4A basal sequence boundary at the top of the Grand Detour (Ludvigson et al., 2004). It thus represents a composited unconformity that encompasses the entire duration of the M4A TST.

The HST of the M4A sequence is interpreted here to initiate at the extremely sharp and distinctive contact between the limestone-dominated, Carimona/Castlewood sub-members and the overlying shale-dominated Glencoe sub-member of the Spechts Ferry Member. Both Witzke and Bunker (1996) and Kolata and colleagues (2001) suggested that the Spechts Ferry was divisible into two "cycles" or "sequences" respectively: the lower, the Carimona (cycle 4A) is a limestone and shale package, and the upper, the Glencoe (cycle 4B) is the shale dominated succession. Kolata et al. suggested that the contact between them was a drowning surface (their DS2). Ludvigson et al. recognized the same contact as a sharp lithologic discontinuity and is developed on top of the Carimona as a bored pyrite encrusted hardground. Farther south the top of the equivalent Castlewood exhibits a prominent phosphatic hardground at the same level. This study interprets this contact as the base of the M4A HST and as it occurs above the Deicke

K-bentonite, it is considered equivalent to the maximum flooding surface observed in Kentucky within the Tyrone Formation and in New York within the Lowville Formation.

The HST succession of the M4A is represented by the lower Glencoe Shale interval. The Glencoe Shale itself is characterized by greenish-gray calcareous shales often interbedded with thin condensed, brachiopod-rich wackestones and packstone beds. The basal shales, below the Millbrig K-bentonite are typically very brown and organic-rich suggesting possible anoxia at the base of the succession. As suggested the sudden loss of carbonates and the appearance of siliciclastic shales strongly suggests a base-level drop accompanied by progradation of siliciclastic sediments – a view that was supported by Ludvigson and colleagues (2004). The Glencoe is absent in northern Illinois and thickens northward toward the Transcontinental Arch where internal marker beds diverge due to increased siliciclastic sedimentation. This suggests that the shales were derived from the latter region (Witzke, 1980). Southward, the shales are occasionally preserved in eastern Missouri and western Illinois, but are thinned or absent at some localities suggesting possible truncation by overlying units.

Upward, the Glencoe can itself be differentiated by internal discontinuity zones (condensed bioclastic beds, ferruginous hardgrounds, and phosphatic nodule beds, etc.), and at least two K-bentonites, including the Millbrig. The Millbrig is intermittently preserved some distance above the brown shales at the base of the Glencoe sub-member, and below the contact with the overlying Garnavillo sub-member of the Guttenberg Limestone. In central eastern Iowa where the entire interval is substantially thinned by non-deposition or sediment starvation, the Millbrig is not recognized and appears to have been truncated by a sequence bounding interval. Thus, as suggested here, the lower Glencoe, below the level of the Millbrig K-bentonite, is likely representative of the entire M4A HST. Unlike in New York and elsewhere on the GACB, substantially little carbonate production occurred in eastern Iowa at this time and siliciclastic accumulation rates were relatively low compared to areas farther north in Minnesota. The absence of the Millbrig and other marker beds in some areas is reminiscent of the situation elsewhere, especially in Kentucky, where the K-bentonite is observed to be truncated below the base of the M4-M5 sequence boundary. Thus here it is suggested that the Glencoe contains the same sequence boundary and the lower Glencoe Shale represents the entire HST of the M4 sequence, although the lower Glencoe may contain at its base an important depositional hiatus of unknown duration.

Cincinnati Arch

The M4 sequence of the Cincinnati Arch is represented by the Oregon through Tyrone Formations of the upper High Bridge Group. The M4A TST is bracketed below by the sharp contact above the Boonesborough shale (Marker Bed I of Stith, 1979) and the sharp contact at the top of the "Herrington Lake shale" recognized here. Thus the dolostones and dolomitic limestones of the Oregon Formation and lower Tyrone Formation are considered to represent the TST and the relatively pure limestones of the upper Tyrone, above the Deicke K-bentonite, are considered to be the HST.

The Oregon Formation, as discussed in chapter 5, is composed of finely crystalline dolostones interbedded with micritic limestones. It is typically a tripartite unit with a lower dolostone, a middle calcilutite, and an upper dolomitic interval. This is amazingly similar to the Nachusa succession of the Upper Mississippi Valley that shows the same tripartite succession. The base of the Oregon is considered to be represented by the sharp facies change out of the silty argillaceous “Teepee”-bearing beds of the Boonesborough shale and is generally a planar to slightly irregular horizon interpreted as the basal SB of the M4 sequence.

The massive-bedded, lower Oregon is characterized by fine to medium-grained dolostones that show repetitive meter-scale cycles that alternate between laminated and bioturbated “leopard-skin” fabrics. The laminated intervals near the caps of the cycles show evidence for algal lamination and mud-cracked intervals and in some cases small, domal stromatolites are also developed. These are succeeded by intraclastic breccias, which in some cases, show clasts of limestones and even small shale intraclasts. Upward into the white-calcilutite-dominated middle Oregon interval, breccias are typically reduced in scale and clast size, but are replaced in small-scale cycles by thin stringers and lenses (ripples?) of coarse-grained skeletal wackestones and packstone beds that continue upward into the overlying dolomitic upper Oregon (and lower Tyrone). The interval is rich in *Tetradium*, crinoids, small twig bryozoans and other fragmented fossils likely influenced by storm activity. Nonetheless, these beds are more fossiliferous than those below, suggesting a deepening upward pattern overall with increased winnowing characteristic of a TST and similar in motif to the lower Lowville that shows the upward loss of dolostones and dolomitic limestones. The upper Oregon dolostone unit is enigmatic, but apparently is a diachronous unit that grades laterally westward across depositional strike from the region of Boonesborough into calcilutite facies interbedded with coarser grained wackestones and packstones of the lower Tyrone Formation. It may be a diagenetic rather than depositional facies. As argued by some workers, the entire formation is not recognized west of the Cincinnati Arch region into the Illinois Basin, where Tyrone-style facies dominate the entire interval. The presence of the dolomitic limestones and dolostones in the Jessamine Dome suggests that the region was more positive early in the M4 TST and less so toward the top of the TST when the lower Tyrone-Upper Oregon facies spread across the area.

The upper TST of the M4 sequence is recorded in the Tyrone Formation which is composed of relatively light, dove-colored, conchoidal fracturing sub-tidal biopelsparites, biomicrites (wackestones), and laminated birdseye micrites that are similar to some portions of the underlying Oregon Formation. Dolostones and dolomitic limestones are rare within the Tyrone. Unlike the underlying Oregon, the Tyrone contains an increased number of clay-rich beds, most of which are thought to be K-bentonites. Portions of the lower to middle Tyrone informally were referred to as the "honeycomb member" after the abundant vertical and sub-vertical *Phytopsis*-style burrows produced in firm micritic sediments. Typically these are infilled with darker-brown, organic-rich micritic infillings which can be dolomitic. Most, however, are not infilled with spar as is characteristic of many equivalent units (including those near the top of the lower Lowville of New York). Nonetheless, these appear to be very similar to those developed in the Quimby's Mill Member of the Platteville Formation of the Upper Mississippi Valley.

At the top of the lower Tyrone is the relatively thick and widespread Deicke K-bentonite that also occurs at the top of the Quimby's Mill Member of the Platteville Formation. As in the Upper Mississippi Valley, the Deicke appears to sit on a discontinuity surface, interpreted as a firm- to hardground surface. Ettensohn and Lowe (2005) observed the contact to contain an abundance of depressed radiating grooves, interpreted as impressions produced by large colonies of the tabulate coral *Tetradium cellulosum* when the substrate was still soft. Subsequently, the surface was reworked by storm activity during a marine inundation and a number of taxa re-colonized the surface as evidenced by the preservation of crinoid holdfasts. The contact is also characteristically set off by a paper-thin, bright-green glauconitic crust that can weather to a rusty color. Here it is considered the equivalent of the base Carimona-top Quimby's Mill discontinuity separating the lower TST from the upper TST. Thus the honeycombed lower Tyrone is likely a condensed interval of stacked firm to hardground surfaces that includes the interval of the Deicke K-bentonite.

Above the Deicke K-bentonite, facies of the Tyrone show an additional interval of meter scale cycles developed in medium-bedded, fine-grained calcilutites that are variously bioturbated and often show small colonies of *Tetradium* corals and occasional wackestone to packstone beds composed of bivalves, ostracods, bryozoans, and brachiopod fragments. Near the top of this succession is a thin, persistent, unnamed K-bentonite that sits about 40 centimeters below a prominent and sharp facies dislocation surface interpreted as the top of the TST and representative of the true MFS.

The interval above the "unnamed K-bentonite" is only slightly coarser, but is composed of dove-brown, intensely-bioturbated, wackestones that grade rapidly upward back into light-gray, fenestral "birdseye" micrites suggesting significant shallowing and is interpreted as the M4HST. Overall, this HST shows substantially more peritidal facies than in the House Creek of New York, but shows similarity with the pattern observed in central Pennsylvania. In the Herrington Lake area on the western side of the Jessamine Dome, and on the eastern side of the dome at Boonesborough, these uppermost peritidal facies show an interval of slightly dolomitic (karstified?), mud-cracked limestones that contain minor stromatolites. These are subsequently capped by the argillaceous beds of the Herrington Lake shale that is correlated with the Weaver Road beds of New York (Brett et al., 2004) and is siliciclastic event five recognized herein. This interval is also likely equivalent to at least a portion of the brown shales of the lower Glencoe

Member below the level of the Millbrig K-bentonite in the Upper Mississippi Valley. The upper M4A sequence is capped by a sharp planar contact at the top of the Herrington Lake shale, above which the “mini sequence” (M4B) is developed immediately below the Curdsville Limestone. This M4B sequence (as discussed below) and the upper Tyrone Herrington Lake shale interval are often locally truncated beneath the more substantial SB at the base of the Curdsville. Thus in some areas the top of the M4A sequence is coincident with the prominent karstified SB influenced by meteoric-groundwater processes, and the M4B “mini sequence” is completely absent. Several meters of incision on this contact, especially in the Frankfort, Kentucky region, suggests this region may have experienced slight to moderate tectonic-induced uplift relative to surrounding parts of the Jessamine Dome at this time.

Central Pennsylvania

In the Ridge and Valley, the M4 sequence is represented by the interval previously referred to as the Curtin Limestone or as the upper Linden Hall Formation. The TST in this region is represented by the bentonite-rich Valley View Member, and the overlying HST is recorded by the Valentine Member of the Linden Hall and its lateral equivalents. The M4B mini sequence is recorded by the Oak Hall Member of the Linden Hall Formation. Collectively the M4 sequence of central Pennsylvania lithologically, is an exceptionally pure limestone with minor bentonitic shales and clay-rich partings.

The basal SB of the M4 sequence in central Pennsylvania is drawn near the top of the Stover Member at the base of the first intraclastic conglomerate and cherty limestone in the lower Linden Hall. The d K-bentonite rests immediately above this bed – which was used by Kay (1944) to separate these two formations. This level is approximately coincident with the top of “mining unit 17” of Laughrey and colleagues (2004) for the Union Furnace region. In the latter region, the basal TST is, in turn, characterized by several meters of relatively pure, thick to massive-bedded, peloidal micrites with occasional interbedded packstones and diffuse chert nodules. These beds grade upward through a number of meter-scale, shallowing-upward cycles that demonstrate overall deepening and become distinguished upward by increased rates of bioturbation in the bottoms of cycles and development of several closely spaced hardgrounds interpreted as the later TST condensed interval. These hardgrounds are often associated with additional intraclastic conglomerate layers – often containing exceptionally pure micrite clasts and larger centimeter-sized, structureless peloids, and bored intraclasts. These are reminiscent of the peloids observed in the underlying sequence during the TST as well. However, the borings in the larger clasts are distinctive here. In many cases, these hardgrounds are complex irregular surfaces and show evidence of burrowing and borings and occasional evidence for corrosion, undercutting, and reworking. Few, however, display significant mineralized contacts, but a few show darkening of micritic matrix in the packstones above the contacts suggesting increased organic matter. This portion of the sequence is also dominated by several K-bentonites that have been recognized previously as occurring within the Valley View Member to the north and northeast. At least two of these K-bentonites sit on hardgrounds and at least one is underlain by appreciable chert. One of these K-bentonites may be the equivalent of the Deicke K-bentonite elsewhere, although this has not been tested. Abundant corals, both tabulate and rugose, are found in this interval as are occasional bryozoan colonies.

Several meters above the K-bentonite interval couplets of brachiopod, bryozoan, and crinoid packstones and grainstones become closely spaced and show evidence of condensation. Similar to the late M4 TST elsewhere, a number of bedding planes at this level, show evidence of extensive, and surprisingly large, *Phytopsis*-style vertical burrows that are often infilled with organic rich micrite and/or chert. Previously, this interval had been interpreted as representing an HST (BR3 sequence of Laughrey et al., 2004); however, the taphofacies and the hardground development in the interval suggests this may be the deepest facies of the TST and it is interpreted herein as representing the cap of the M4 TST and the maximum flooding. As elsewhere, this contact is represented by a sharp discontinuity and facies dislocation surface, but here is marked by the change out of packstone lithologies and into massively-bedded, bioturbated mudstones and wackestones, that do in this case show some minor silty argillaceous partings immediately above the B5 K-bentonite.

In the Valley View type area, the TST facies is more similar to that of the lower Lowville of New York, in that the packstones and grainstones developed in the Union Furnace region are not represented in the TST. Instead in the region near Bellefonte, the Valley View is dominated by a succession of slightly fossiliferous *Tetradium*-bearing, peritidal peloidal micrite facies with chert-rich horizons and slightly shaly (bentonitic?) partings between beds. This unit overall does not show significant lithologic change upward to the MFS, except for slightly coarser beds and slightly shalier beds near the contact with the overlying Valentine Member of the Linden Hall. In this region, only minor evidence exists for hardground development, although the K-bentonites do still appear to be developed on slightly irregular contacts. This region was evidently a much shallower area throughout the TST, and may have been slightly more restricted.

To the south, in the Union Furnace area, the massive bioturbated wackestone facies of the lower HST change upward across a prominent intraclastic (channelized?) grainstone bed into a highly cyclical interval composed of interbedded, thin to medium-bedded fine calcilutites and slightly coarser-grained dolomitic wackestones of the Oak Hall Member of the Linden Hall. In many cases, these beds show evidence for rippled-bedding, occasional cross-bedding, and an increase in ctenodont bivalves and ostracods, bifurcating burrows, and rare, but highly abraded brachiopod hash beds. This facies also contains occasional rugose corals and fragmented *Tetradium*. Overall, this facies suggests a shallowing-upward pattern, albeit not back to fully peritidal facies at this locality. In the Bellefonte region and a few other localities, this same interval is characterized by a massive conchoidal-fracturing micrite unit that is nearly structureless and devoid of fossil material. In a few cases, rare *Tetradium* and ostracods are recognized and possible fenestrae or birdseye structures may be present as well as occasional cryptalgal laminations. The unit is very peculiar and is not a typical facies, but it can be traced laterally into the lower Oak Hall which enables its recognition as a regressive deposit of the HST. Moreover, its top contact is irregular and complex owing to erosional truncation, incision, and even possible karstification. The unit is clearly truncated to the east of the Nittany Valley onto the Adirondack Arch and is overstepped by the overlying Nealmont Formation. In the Bellefonte region, Kay (1944) recognized channels cut down into the unit and infilled with coarse-grained facies of his Oak Hall Member of the Nealmont Formation (upper Oak Hall of Rones, 1969). Here this Oak Hall facies (upper Oak Hall of Rones) is interpreted as the latest HST/RST deposit and the coarse-grained facies, infilling the channels as representative of the M4B "mini sequence" that leads up to the M4-M5 SB at the top of the Oak Hall Member (see below).

M4B "Mini Sequence"

As discussed above, the top of the M4 sequence contains a relatively thin interval (typically no more than two meters in thickness) and is similar in scale to some parasequences (**figure 13**). As such, it might appear to represent the RST to LST deposit of the M4 sequence. It is not present in all localities and is often removed by the subsequent SB. Nonetheless it is a

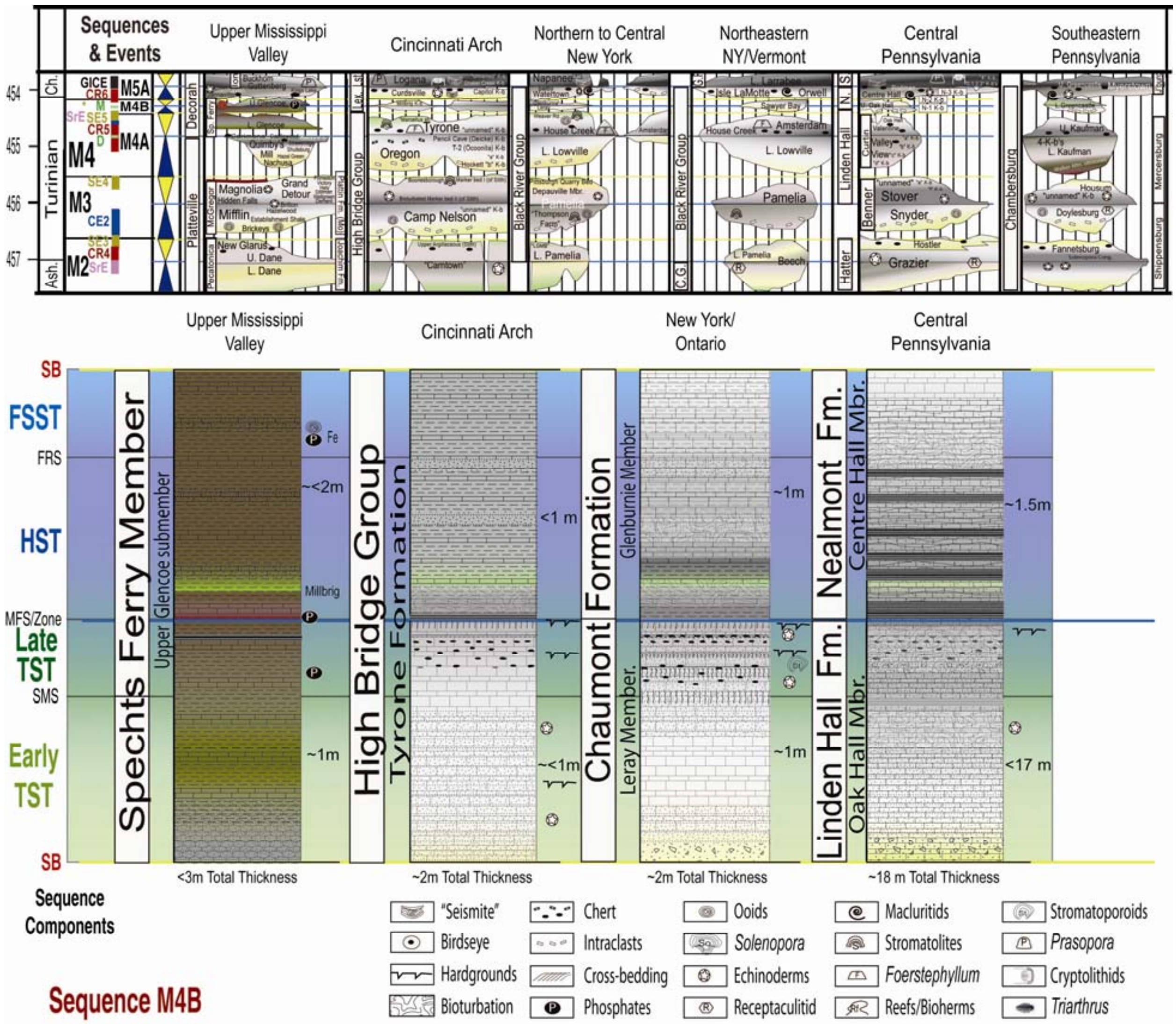


Figure 13: Correlated framework of the M4B "mini sequence" (Turinian) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania. As the sequence contains the Millbrig K-bentonite, this sequence contains the Turinian-Chatfieldian chronostratigraphic boundary.

distinct unit that shows a substantial magnitude of facies change and its own distinct TST, MFS, and HST components that are typically not developed in underlying sequences. Thus it behaves as a sequence, although it is very thin and is sandwiched between two much larger and thicker sequences.

In New York State, the M4B sequence is represented in the interval above the Weaver Road Beds (siliciclastic event 5) and the base of the Watertown Limestone and its basal SB. The intervening interval is thus characterized by the Leray Limestone and the Glenburnie argillaceous limestone and shale. Moreover, and most importantly, this interval contains the Hounsfield K-bentonite (sensu Kay, 1931) which is now correlated with the Millbrig K-bentonite (see chapter 2 for discussion). In the Lake Simcoe region of Ontario, a subtle facies dislocation occurs at the base of the lower Coboconk Member of the Bobcaygeon Formation. Described in detail by Okulitch (1931), this horizon appears as a sharp transition out of bioturbated *Tetradium*-rich, slightly argillaceous, stromatolite-bearing, and rubbly weathering wackestones and mud-cracked micrites of the uppermost Moore Hill Member (House Creek – Weaver Road Bed equivalents) into stromatoporoid and coral bearing, cross-bedded, crinoidal grainstones of the lower Coboconk Member. These latter facies are clearly higher-energy deposits and likely represent shoaling-upward transgressive conditions out of the low-energy, peritidal facies below.

In New York, the contact also places coral-bearing crinoid grainstones abruptly over shaly micrites and platy shales of the Weaver Road member. This is a sharp, slightly wavy surface that locally truncates the upper Weaver Road beds. Occasional rip up clasts of the underlying units are found within the basal coarse skeletal hash bed, especially in sections near Watertown and Lowville, New York. Because there is some evidence for local erosion at this contact it is considered to be a combined FRS to lowstand erosion surface (SB) and transgressive ravinement surface (TS). The coarse-grained unit above the contact often contains silicified fossils and/or quartz-rich sandy zones at its base. These beds clean-upward to a relatively pure coquinal rudstone cap and are thus interpreted as a small-scale TST during a relatively rapid period of base-level rise. The TST is typically easily distinguished from underlying beds as it contains a more diverse fauna including gastropods, brachiopods, crinoids, corals, and bivalve fragments. The top of the bed is a sharp planar contact (MFS), above which is an interval (less than one meter-thick) of shaly nodular limestones to platy shales (often organic rich and yellow weathering).

These shales, referred to as the Glenburnie Shale (Kay, 1931), carry a diverse fauna including brachiopods and bryozoans many of which are known from the Decorah Formation of the upper Mississippi Valley. These beds are interpreted here to reflect an influx of siliciclastic sediments during the HST of the M4B mini sequence, although they are relatively thin. This interval also contains the Hounsfield/Millbrig K-bentonite. These shales and argillaceous limestones are only recognized in the vicinity of the Kingston Trough (NY & Ontario) and are shown to pinchout onto the Marmora and Adirondack Arches and are truncated below the overlying Upper Coboconk/Watertown Formation and the basal SB of the M5A sequence.

CORRELATION OF THE M4B MINI SEQUENCE

Upper Mississippi Valley

In the upper Mississippi Valley, the M4B mini sequence, like in New York and Ontario, appears to be truncated in northern Illinois and in portions of Iowa where the TR-4, cycle B (M5A of this report) of Witzke and Bunker is shown to remove the underlying TR-4, cycle A (M4B, herein). Northward, into northeastern Iowa and southern Minnesota, the M4B sequence is recorded within the Glencoe Shale that also contains the Millbrig K-bentonite. Overall the Glencoe is an extremely starved facies dominated by shales and occasional packstone coquinal layers, and abundant phosphatic stained surfaces and nodule-bearing horizons, and even iron-ooids (Emerson, 2002). Where the Millbrig is preserved, there is some evidence for possible

M4B TST as represented by thin coquinal lag beds composed of bryozoans and brachiopods underlying the K-bentonite. These grade back to greenish-gray shales and interbedded siltstones above the Millbrig and all below the level of the Guttenberg Limestone. As mentioned elsewhere, most of these shales and siltstones thicken into the Hollandale Embayment to the north and appear to have derived from the Transcontinental Arch (Emerson, 2002). These likely represent progradation during the latest HST/RST of the M4 sequence. The phosphatic and iron-ooid rich upper Glencoe of the Hollandale Embayment in Minnesota likely records the LST and or condensed interval going into M5A sequence – the latter of which is recorded by the return to clean carbonates at the base of the Guttenberg Formation.

Cincinnati Arch

The M4B sequence in the Jessamine Dome region is more similar in construct to the succession developed in New York, than that of the upper Mississippi Valley. In Kentucky, the interval immediately above the green Herrington Lake shale shows a distinct change to pinkish, fine- to medium-brachiopod and crinoidal grainstones, with a basal silty limestone. This relatively pure, grainstone unit (~1 meter thick) is capped by a thin interval of three or four greenish-gray, wavy-bedded calcisiltite beds that show minor quartz silt and slightly argillaceous components. Lying at the top of these greenish beds is the Millbrig K-bentonite, which in some localities is erosionally truncated in the face of the outcrop below the overlying coarse-grained limestones of the Curdsville Member. The basal grainstone interval is thus interpreted as the TST, the silty beds with the Millbrig K-bentonite is interpreted as the HST, and the sharp, planar contact between these facies is interpreted as the MFS of the mini sequence.

In the Herrington Lake region near Marcellus, Kentucky, this interval shows evidence that the succession is really composed of two meter-scale parasequences that are stacked together to form the mini sequence. However in other localities, especially where the Millbrig has been truncated by the overlying M4-M5 SB, the coarser grainstones of the Curdsville become superimposed on the fine- to medium-grained grainstones of the uppermost Tyrone above the Herrington Lake shale and are thus all referred to as Curdsville, even though the SB may actually occur just above the first grainstones in some localities. In other localities, it appears that the entire M4B and a significant portion of the upper M4A sequence is erosionally truncated by the prominent M4-M5 SB.

Central Pennsylvania

In the Ridge and Valley, the M4B mini sequence is represented by the upper Oak Hall Member of the Linden Hall Formation to lowermost Centre Hall Member of the Nealmont Formation. The unit shows a much more variable thickness than its equivalents in New York or Kentucky and a wider range of lithologies. The base of the upper Oak Hall, a fine- to medium-grained, cross-bedded grainstone unit, was shown by Kay (1944) to channel into the Valentine Member in some localities, with the coarser, ledge-forming grainstones of the upper Oak Hall Member forming the channel-fill deposits. Here these channel features may represent the late HST/FRS surface and the RST is represented by the grainstones. This portion of the Oak Hall grades upward into darker gray, medium- to coarse-grained packstone facies into the base of the Centre Hall Member.

The Centre Hall is substantially thinner-bedded and much more argillaceous (ribbon facies) before it is capped by massive bioturbated wackestones and cherty packstone facies of the upper Centre Hall. This succession thus appears to record an upward-deepening pattern up to the sharp facies dislocation surface about 1.5 meters above the N1 K-bentonite. As in New York this TST succession is typically fossiliferous with abundant corals, brachiopods, and other fossil debris. The argillaceous ribbon, to slightly nodular, facies of the overlying Centre Hall contains the N2 K-bentonite which is postulated to be the equivalent of the true Millbrig K-bentonite. The sudden appearance of the argillaceous beds here is coincident with the HST of the M4B mini sequence and is thought to record base-level drop and the progradation of siliciclastics during the last part of the M4 sequence. As in the Upper Mississippi Valley, there is a slightly greater thickness of strata preserved above the suspect Millbrig compared to either the Cincinnati Arch area or the New York platform region. The increased thickness in this region, and the lack of a major erosional unconformity in the Centre Hall above this level, strongly suggests this region was beginning to founder and subside rapidly at the onset of the M5 sequence. Nonetheless, for a short period, base-level rose again and the upper Centre Hall Member becomes less argillaceous during the M5 TST.

M5-C1 SEQUENCES OF THE "TACONIC SUPERSEQUENCE"

The M5 to C1 sequences coincide with the uppermost Black River Group through the Trenton and into the Utica Shale of New York. In Kentucky and Ohio, the M5 through C1 coincide with the Lexington and Point Pleasant Limestones and the overlying Kope Formation. In central Pennsylvania, the interval includes the Nealmont, Salona, and Coburn Formations respectively. The M5 and M6 sequences have been subdivided here, each into three sub-sequences that have been previously outlined for New York and Ohio (see Brett et al., 2004). The following expands the discussion to include correlation of these sequences into the upper Mississippi Valley and central Pennsylvania.

M5A Sequence: Curdsville-Logana; Watertown-Selby-Napanee; Nealmont-New Enterprise

In the Nashville Dome, the M5A sequence is composed of the Curdsville Limestone (M5A TST) and the lower Hermitage Formation (M5A HST) of the Nashville Group. The sequence is very important in that it contains several prominent K-bentonites above the level of the Millbrig, and in Tennessee the sequence contains the *Phragmodus undatus-Plectodina tenuis* chronozone boundary (Leslie, 2000) and records a significant extinction in both trilobites and cephalopod faunas as discussed in chapter 2. The sequence also contains the Guttenberg Isotopic Carbon Excursion (GICE), chert-rich interval 6 and other chemostratigraphic events documented in chapters 2 and 6. Moreover, the sequence also records the first major pulse of siliciclastic sedimentation that inundated the craton clear across to the upper Mississippi Valley with sediments derived from the newly uplifted Taconic highlands. The important trigger, of which, was the subsidence of the Adirondack Arch as a prominent topographic feature. This resulted in modified circulation patterns in the GACB during a substantial sea-level rise event, and perhaps associated with a major climatic shift at this time.

In New York and Ontario the M5A sequence (**figure 14**) is bounded below by the SB recognized by Cornell (2001) at the base of the Watertown Member of the Chaumont Formation. The basal boundary is a sharp, nearly planar or very gently undulatory contact between the massive grainstones, and packstones of the Watertown and underlying units discussed previously. The beds above the SB can contain rip-up clasts derived from immediately subjacent beds. As the Watertown grainstone facies is traced southward, its basal contact oversteps (and truncates) underlying units and eventually transitions laterally into massive bioturbated wackestones and eventually into thin birdseye limestone facies that are often inseparable from

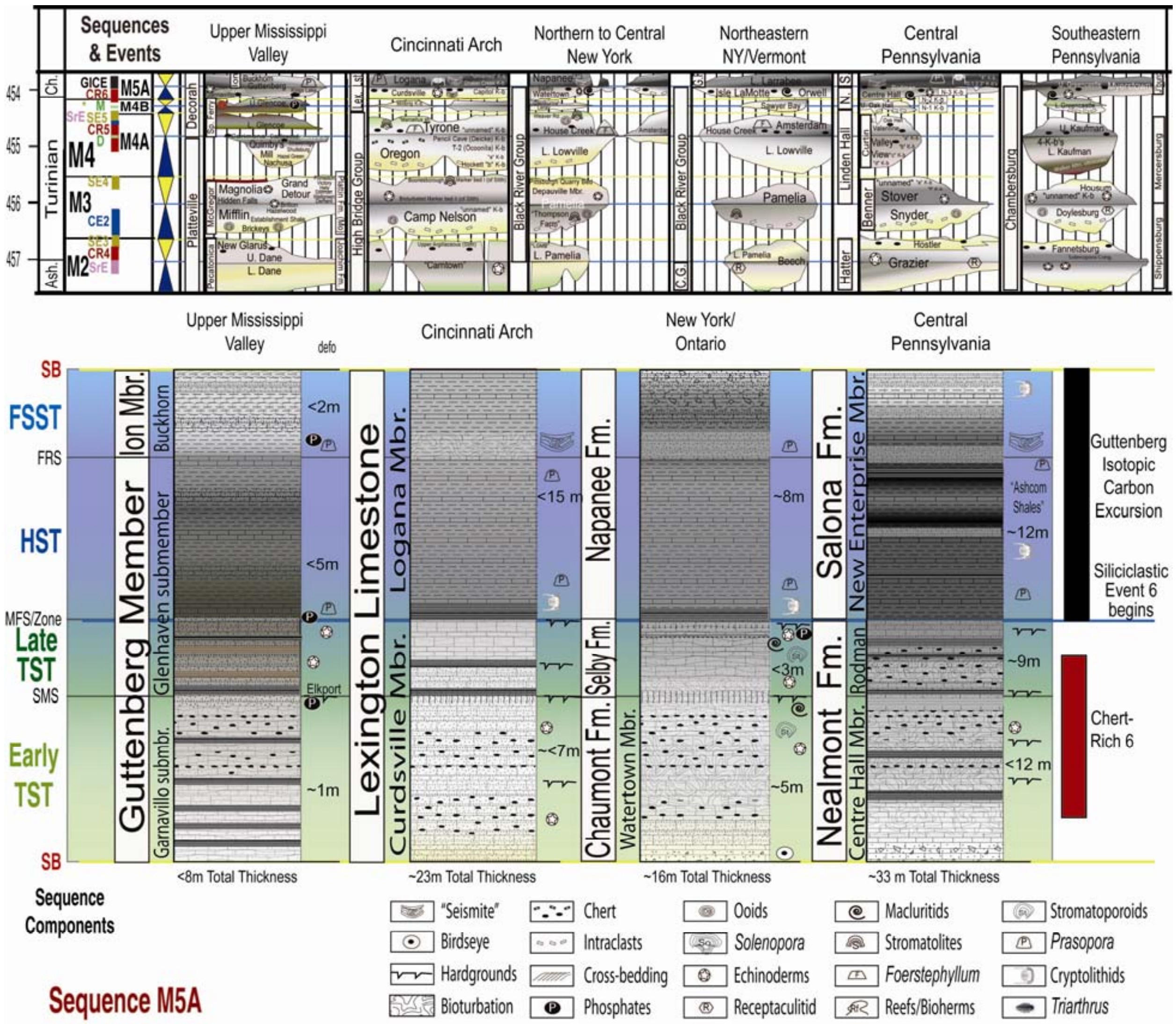


Figure 14: Correlated framework of the M5A sequence (the first Early Chatfieldian sequence (~Rocklandian)) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

the underlying Lowville Formation without recognition of the sequence boundary. However, near Newport and Middleville, New York (eastern Mohawk Valley) the Watertown rests directly on the highly cyclic beds of the House Creek Member of the Lowville Formation. The same erosion surface in this more proximal section along the Adirondack Arch is developed into an irregularly channeled, and karstified, erosion surface which removed about half of the House Creek Member. At one locality there is evidence even for cave development below the contact surface that subsequently collapsed into a sinkhole during deposition of the M5A HST. Because of the strong evidence for regional beveling of units below this contact, it is regarded as a major sequence boundary and is discussed in more detail elsewhere (see Cornell, 2001; Brett et al., 2004).

The TST of the M5A sequence is marked by massive, locally cherty, bioturbated skeletal wackestones and packstones of the Watertown Formation and overlying Selby Formation. The contact between the Selby and the underlying Watertown is fairly sharp and shows evidence for condensation and sediment starvation especially in the southern Black River to western Mohawk Valley regions. The upper portion of the Watertown and the Selby, both contain, mineralized hardgrounds, which in some cases contain K-bentonites that have been fingerprinted and correlated with some issue as discussed in chapter 2 and 6 of this report. In the Mohawk Valley, the Selby is most extensively condensed down to a single vertically burrowed/bored, conchoidal-fracturing calcilutite facies. The burrows in this unit are infilled with black, organic-rich matrix, considered to be anthraxolite that carries heavily fractionated carbon isotopic values (Argast, 1992). Thus from base to top, the Watertown to Selby interval shows evidence of deepening and in some cases extreme condensation especially in the Mohawk Valley, where it appears that the region suddenly subsided from peritidal conditions to significantly deeper-water environments coincident with sea-level rise. As such the Watertown likely represents the early TST and the contact between the Selby and the underlying Watertown probably represents a sediment starvation surface at the base of the condensed interval (late TST).

Across the region, the MFS is represented by the sudden shift from relatively pure limestones into siliciclastic-influenced carbonates of the Trenton Group. In the central Mohawk Valley, the shaly calcisiltites of the Napanee Formation overlie the Selby-Watertown interval with pronounced discontinuity. This sharp corrosion surface was formerly interpreted as the Black River - Trenton SB. Cornell (2001) reinterpreted this contact as a submarine corrosion surface, with evidence for dissolution, pyritization, and a high gamma-ray signature typical of MFS condensed intervals. Thus, this surface is now considered to be the MFS of the M5A sequence.

The HST interval of the M5A sequence is represented by the Napanee Formation of New York -Ontario. This interval is dominated by rhythmically interbedded calcilutite/calcisiltite and dark shale facies, and commonly shows a thick, amalgamated, middle bed of dalmanellid-rich packstones and occasionally grainstones, that may represent a portion of a minor 4^{th}-order sequence. In Ontario, the Napanee appears to correlate with the first thin-bedded and shaly facies of the middle Bobcaygeon Formation above the upper Coboconk (the TST of the sequence). Taphonomic as well as biostratigraphic evidence from acritarchs and chitinozoans indicate that the Napanee is the deepest facies of the sequence. Moreover, the unit generally shows an aggradational to slightly progradational shallowing-upward pattern toward the contact with the overlying Kings Falls Formation and is characteristic of a HST. This is the facies that records the positive isotopic carbon values characteristic of the GICE event (Barta, 2004).

In the southern Black River to eastern Mohawk River Valleys, there is evidence of a sharp discontinuity surface and development of a few coarser-grained skeletal packstones near the very top of the Napanee Formation. These beds continue to shallow upward, contain fragments of *Tetradium* corals, and even contain occasional small lithoclasts of underlying units (and even small quartz pebbles). These, in turn, are sharply overlain by a thick succession of very coarse-grained calcarenites of the Kings Falls Formation with substantially larger clasts, some of which appear to be derived from the Grenville-aged basement. The basal contact of the cross-bedded and rippled calcarenites of the Kings Falls represents the SB at the top of the M5A and the upper Napanee *Tetradium*-bearing shallowing succession below them is inferred to represent a very thin late HST to RST interval. Farther to the northeast, these late HST deposits are not easily identified and may not, in fact, be developed below the M5B basal SB.

CORRELATION OF THE M5A SEQUENCE

Upper Mississippi Valley

Given that this is the type region of the GICE event, that is recorded in Kentucky, New York, and elsewhere, this sequence is critical to establish correlation into the upper Mississippi Valley. Here, the M5A sequence is correlated roughly with a transgressive-regressive cycle (TR cycle 4B) recognized by Witzke and Bunker (1996). These authors recognized that the Spechts Ferry Shale represented a significant progradation of muds prior to the deposition of the Guttenberg Limestone. The return to more abundant carbonate deposition was interpreted as a renewed transgression and initiation of their cycle 4B. The lower portion of the sequence, the SB to TST interval is represented by the condensed, wavy- to nodular-bedded limestone strata and phosphatic lag deposit of the basal Guttenberg (Garnavillo sub-member). These beds resemble the McGregor Member below the Spechts Ferry and are often very fossiliferous. The Elkport K-bentonite occurs at the top of the Garnavillo sub-member and may sit in the position of the major flooding surface going into the late TST. Witzke and Bunker (1996) suggested that the lower Guttenberg recorded a deepening-upward pattern with whole-shell skeletal mudstone fabrics and few coquinal layers indicative of deposition below storm wave base. Although the absence of coquinal layers and high shale content suggests the deepest water facies, it may also represent relatively shallow water, low-energy deposition in a protected (restricted?) embayment located behind a restrictive barrier. Here it is suggested that this lower Garnavillo facies, does record deepening, but may only record early TST base-level rise that continues upward into the middle to upper Guttenberg.

Evidence supplied by Witzke and Bunker (1996) and Ludvigson and colleagues (2004) suggests an upward increase in tempestites and amalgamation of limestone beds into the middle to upper Guttenberg (Glenhaven sub-member). This was interpreted by these authors as suggesting depositional shallowing and the regressive phase of the cycle (HST). However, as elsewhere, this facies, with more abundant coquinal layers, including many that contain abundant echinoderm faunas (see Kolata, 1975), may represent evidence for upward shoaling conditions and increased winnowing into the late TST. Moreover, this facies is also known for its brown organic-rich shale partings (with up to 50% total organic carbon suggesting significant condensation) and thin interbeds of green-gray shales which increase upward in the transition into the Ion Member. This upper Guttenberg may thus actually represent the deepest facies of the sequence and record initiation of base-level drop during the early HST. Also interesting is the fact that these middle to upper Guttenberg shales carry not only the signature of the GICE, but also neodymium isotopic values characteristic of Taconic-derived sediments for the first time. Bedding styles in the upper Guttenberg become less wavy-nodular and more planar bedded and then grade into significant amounts of greenish gray calcareous shales interbedded with fossiliferous wackestones and packstones including large *Prasopora* bryozoans. The regional progradation of this packstone and shale-dominated facies is interpreted as the later HST interval with the rapid increase in shale in the Ion Member (Buckhorn sub-member) of the Decorah Formation representing late progradation associated with the RST. It is also worth noting that these shales show a return to Transcontinental Arch-derived neodymium values indicating local transport of sediments once again as Taconic-derived sediments became limited.

As interpreted by Fanton and Holmden (2007) locally-derived neodymium (from the Transcontinental Arch?) swamped the signal from the east, although accentuation of portions of the Adirondack Arch during the late M5A sequence may have helped to once again reduce the transportability of siliciclastics to the craton interior. This strongly suggests this to be the M5A-M5B sequence boundary interval.

Cincinnati Arch

In the Cincinnati Arch, the M5A sequence boundary is represented by the sharp basal contact of the Curdsville Member of the Lexington Limestone. As with the base of the Watertown Limestone in New York, and as discussed above, the basal Curdsville unconformity shows evidence of significant truncation across portions of the Jessamine Dome and is more pronounced in some areas interpreted to have been uplifted locally (specific details are outlined by Brett et al., 2004).

As discussed in chapter 5, the Curdsville is composed of three sub-units. The lowermost is a fine-to-coarse grained, cross-bedded, and ripple-marked grainstone and calcirudite with, occasional interbeds of laminated calcisiltites and up to ten percent fine quartz sand. Most skeletal grains in this unit are highly abraded and sorted fragments of brachiopods, crinoids, bryozoans, and bivalves. The middle Curdsville is another bioclastic calcarenite and calcirudite-bearing unit, but in this case these facies are interbedded with argillaceous calcisiltites and dark carbonaceous shales at about the base of the Capitol metabentonite. This interval contains minor chalky-white cherts and rippled-calcarenites that are commonly developed into hardgrounds with preserved encrusting organisms attached to them. The uppermost unit of the Curdsville Member is an irregularly bedded, bioclastic, fine-grained grainstone that is typically interbedded with closely spaced brachiopod coquinas that show excellent preservation.

Thus the Curdsville represents an overall deepening upward succession and is the TST of the M5A sequence. The TST shows excellent development of an early, middle, and late TST interval the middle of which often contains occasionally well-preserved echinoderm faunas similar to those of the Guttenberg of the upper Mississippi Valley, and the upper Coboconk and Kirkfield of Ontario. The middle Curdsville condensed interval is also typically quite chert-rich and is identified here to correlate with chert-rich interval 6.

This facies is followed by a significant change to planar-bedded calcisiltite “rhythmite” facies of the Logana Member of the Lexington. The limestones are interbedded with shales that makeup up to 50% of the unit. Calcisiltites contain relatively few articulated fossils – and commonly exhibit evidence for storm-influenced deposition. The dark shales are often carbon-rich and impart a petroliferous odor. In this case, the occurrence of *Cryptolithus* trilobites in the base of the Logana, coincident with the darkest shales of the interval, indicates that this unit represents the deepest facies of the M5A sequence and is as such interpreted as the early HST. Upward the center sub-unit of the Logana is dominated by coquinal packstones and grainstones interbedded with calcisiltites and shales. This middle Logana package likely represents the transition to the late HST. In some cases, the middle unit contains evidence for soft-sediment “seismite” deformation which may have formed in response to rapid deposition during the later HST. The top of the M5A sequence is drawn at the sharp contact with the Grier Member. The contact juxtaposes medium- to coarse-grained packstones and calcarenites above the deeper calcisiltites of the Logana. This boundary is hence considered the SB (see below).

Central Pennsylvania

In the Ridge and Valley region of central Pennsylvania, the M5A sequence is here interpreted to be the middle to upper Nealmont Formation (Centre Hall and Rodman Members) and the lowermost New Enterprise Member of the Salona Formation. The Nealmont Formation records the last massive, and relatively pure carbonate unit before rapid subsidence and major shale influx into the region. As discussed the lowest part of the Centre Hall is somewhat argillaceous and was interpreted as the HST of the M4B mini sequence. The upper part of the Centre Hall and overlying Rodman Member begin to show the change out of relatively shallow water, bioturbated, and *Tetradium*-bearing calcilutites back to crinoid, brachiopod, and bryozoan-dominated calcarenites. This latter facies is spread over much of the Ridge and Valley region and oversteps many underlying units against the Adirondack Arch so that the Nealmont rests on top of much older M3 and even M2 sequence strata on the Adirondack Arch before it transitions into facies of the upper Greencastle and Myerstown Formations of the Great Valley area. Thus, similar to the Curdsville Limestone of Kentucky, this facies is a relatively homogeneous, sheet-type deposit across its area of distribution and is easily recognized in northern Virginia and West Virginia.

Faunally, the Centre Hall is a coral, gastropod, brachiopod-dominated succession that contains unique species in common with the Watertown Limestone of New York, the Guttenberg of the upper Mississippi Valley and the Curdsville Limestone. These slightly argillaceous facies grade upward into much darker, organic-rich limestones of the Rodman bearing crinoids and other echinoderms as well as bryozoan thickets, with only one coral species listed in the lower Rodman (Kay, 1944). Although the Rodman contains a number of coquinal beds, preservation of some echinoderms is occasionally good suggesting intermittent depositional energies under deeper water conditions. Combined with the Centre Hall, this unit is interpreted as the TST of the M5A sequence in this region. The Rodman shows a pattern of upward thinning of beds and a change from more massive bioturbated wackestones into more nodular, argillaceous calcarenites interpreted as the late TST interval. The interval also is typically cherty and thus characteristic of another chert-rich interval (recognized as chert-rich interval 6).

The MFS of the M5A sequence is represented by the facies dislocation from the Rodman into the overlying Salona Formation. The contact is often sharp and planar and shows the juxtaposition of the first thin- to medium-bedded and often structureless rhythmically-bedded, tabular calcilutites and calcareous shales over the condensed fine-grained grainstones of the Rodman Member. Like the underlying Nealmont, this interval of the Salona (the lowermost New Enterprise Member) is the most widespread and most uniform of facies in the western Ridge and Valley. The characteristic calcilutite-calcisiltite facies is peculiar in that it is nearly structureless and only shows minor evidence for basal scour followed by normal graded bedding with occasional skeletal lag beds flooring the deposit. These are interpreted as distal storm beds and suggest very deep-water conditions that deepened slightly to the south of central Pennsylvania toward the Maryland line. Some of the more prominent, and most organic-rich shales, occur some distance above the base of the Salona and contain common graptolites and abundant *Cryptolithus* trilobites. The presence of these shales (informally referred to as the "Ashcom shales") some distance above the Rodman-New Enterprise contact suggest this may actually record the maximum flooding zone and the change to the early HST of the M5A sequence.

Upward above the lower Salona K-bentonites, facies show a decrease in overall bed-thickness and an increase in grain size with the occasional occurrence of fine-grained grainstones with minor hummocky cross-stratification near the middle of the New Enterprise Member, suggesting some shallowing. Shales are still abundant and occasional silty beds interfinger from the south and east, although these are typically very minor at this time. As this unit is traced into the southeastern-most valleys of the Ridge and Valley, it grades rapidly into the lower carbonate-rich shales and siltstones of the Martinsburg Formation.

M5B Sequence: Kings Falls- Sugar River; Grier; Upper New Enterprise

The sharp juxtaposition of fossil rich cross-bedded, crinoidal grainstones over the characteristic Napanee lithology signals the SB between the M5A and M5B sequences (**figure 15**). In the Mohawk Valley, and occasionally elsewhere, beds above this sharp contact contain

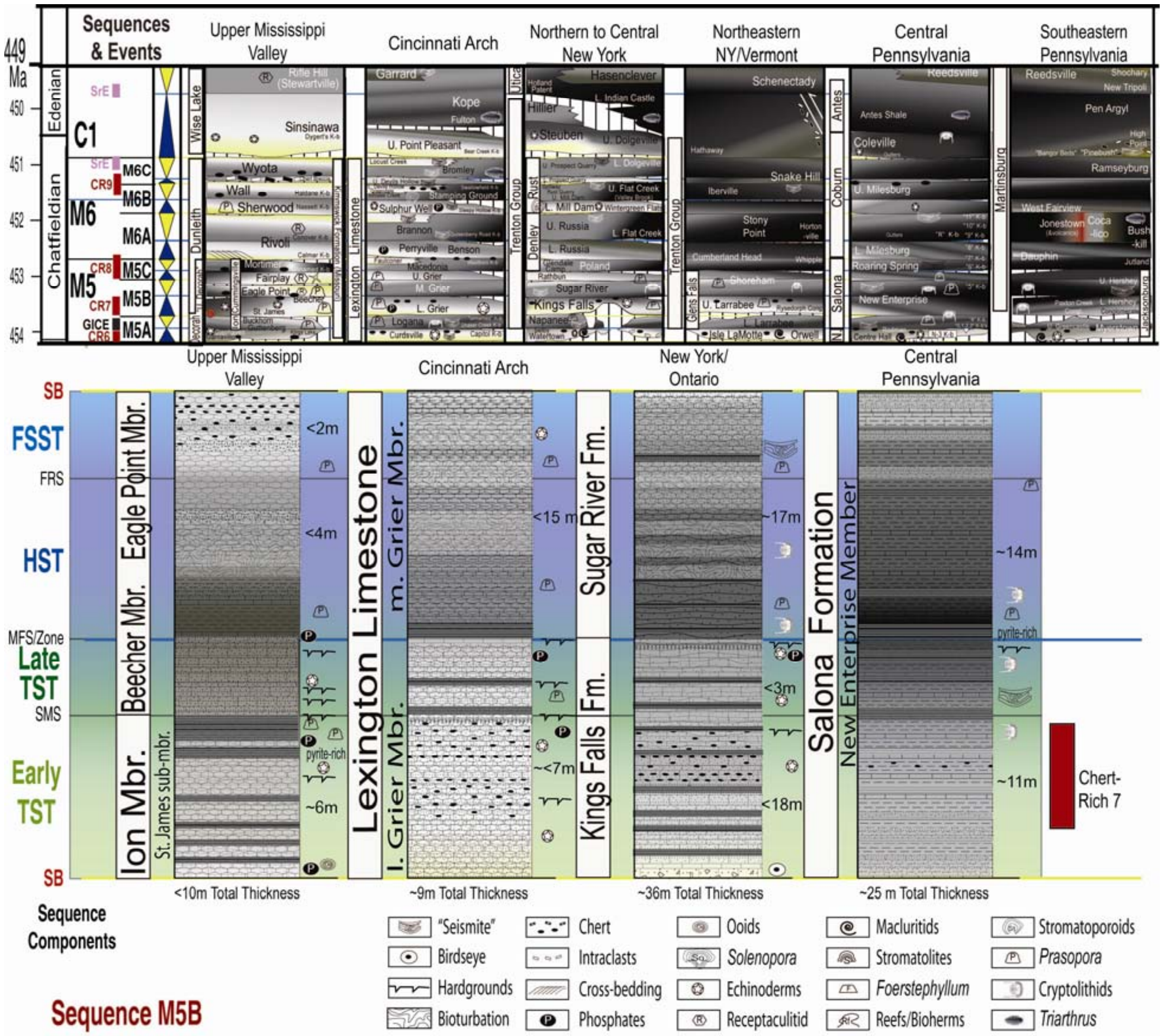


Figure 15: Correlated framework of the M5B sequence (Mid-Chatfieldian sequence (Kirkfieldian-Shermanian Boundary)) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

pebbles of the underlying Napanee Formation. The unconformity may also cut down through the Napanee to merge with the M4B-M5 SB especially where the Kings Falls rests directly on mud-cracked micrites of the Watertown equivalent or the Lowville Formation (Cornell, 2001). Irregularities in the extent of truncation and in the origin of lithoclasts within the basal Kings Falls, indicates that the Mohawk Valley region became substantially modified by local tectonic activity at this time (Bradley & Kidd, 1991). Evidently numerous fault-bounded blocks began to move coincident with deposition of the Napanee and Kings Falls formations, both in New York and in Pennsylvania. Farther east, across the Canajoharie Arch, the grainstone-rich interval of the Kings Falls grades laterally into the upper Larrabee Member of the Glens Falls Formation. In this easternmost area, the unit begins to show significant evidence for condensation of grainstone intervals and a rapid increase in the relative proportion of dark shales deposited in the advancing Taconic foredeep basin. In the Lake Simcoe area of Ontario, a correlative surface lies at the sharp base of the upper Kirkfield (upper Bobcaygeon) grainstones and rests sharply on the underlying lower Kirkfield (middle Bobcaygeon) shaly beds (see Cornell, 2001). It is not clear if there is erosion at this contact, but it does represent a sharp facies dislocation and a change from shallowing-up to deepening-up facies.

In central to northern New York, the Kings Falls Formation shows evidence for rapid upward deepening with very coarse grained facies grading upward into more condensed shaly nodular brachiopod, and echinoderm packstones and wackestones. The Kings Falls represents both the early and later condensed portions of the TST, and in Ontario, the upper portion of this unit contains well-preserved echinoderm faunas characteristic of late TST to early HST deposits in Lexington Limestone sequences. The top of the M5B TST is slightly obscure due to the lack of outcrop exposures. However, in the southern Black River Valley region, the top of the Kings

Falls shows evidence for condensation and phosphatization of skeletal grains before transitioning into Sugar River shaly nodular wacke- to packstones with abundant *Prasopora* seen in much of central New York. The maximum flooding surface of the M5B sequence is placed at the contact of the Kings Falls and the Sugar River formations. In outcrop exposures of the type Kings Falls Formation this contact is just below a distinctive irregularly-bedded disturbed zone within the overlying Sugar River HST.

The M5B HST is composed of shaly wavy bedded pack- to fine-grained grainstone facies and burrow-nodular wackestones. Like the M5A HST, the latter facies are often noted for beds containing *Prasopora*, and *Cryptolithus tesselatus*. This lithofacies represents deposition in substantially deeper water setting than much of the underlying Kings Falls, although not apparently as deep as the Napanee Formation. The lower Sugar River Formation is generally aggradational in character, but shows some minor evidence for progradation and rapid deposition of some storm beds including occasional *Prasopora* conglomerates and the first soft-sediment deformed intervals recorded in New York. In eastern New York, the highstand facies of the Sugar River sequence shows an abrupt upward change to dark Flat Creek Shales, the base of which shows evidence of extreme starvation, such as phosphatic-pyritic staining, and is inferred to be the result of tectonically enhanced deepening associated with the migration of the Taconic Foreland Basin into eastern New York.

Near the contact with the Rathbun Member of the Sugar River, the transition into the later HST is associated with increased bioturbation and subsequently a change to relatively pure calcilutite facies with condensed skeletal lags. Although dominantly fine-grained, this facies is distinctive in the loss of substantial shaly interbeds and signifies a renewed period of base-level rise associated with the onset of the M5C transgression.

CORRELATION OF THE M5B SEQUENCE

Upper Mississippi Valley

Witzke and Bunker (1996) recognized their transgressive-regressive cycle T-R cycle 5A as a "lumped" cycle composed of strata of the lower Dunleith Formation. This is equivalent to the Kimmswick Limestone of Missouri, and roughly to the lower half of the Cummingsville Formation of the Galena Group in Minnesota. Overall, this sequence is dominated by argillaceous, fine-grained, and thin to occasionally nodular-bedded limestones that are typically interbedded with conspicuous thin to thick shaly partings and bearing abundant *Prasopora* (Weiss, 1955). This unit is essentially equivalent to what some authors refer to as the "Upper Decorah" of southeastern Minnesota and northeastern Iowa (Witzke and Bunker, 1996). It is a facies very similar to much of the Sugar River Formation of New York State and the Grier Member of the Lexington Limestone of Kentucky with which it is correlated here.

TR- cycle 5A is shown to onlap strata around the Ozark Dome in Minnesota and Illinois and portions of the Wisconsin Dome. As suggested by the former authors, the change from shale-dominated facies of the Buckhorn sub-member of the Ion to carbonate-dominated facies of the St. James sub-member of the Ion Member indicates general depositional deepening above the M5A sequence and therefore initiation of the M5B TST. Within the shalier facies of the Cummingsville (in Minnesota) / St. James sub-member of the upper Decorah (in Iowa) the SB at the base of the M5B TST is recorded by oolitic ironstones and reworked phosphatic nodule facies similar to those recognized in the late part of the M4B mini sequence below the M5A TST. Neodymium isotopic values at the base of the TR-5A show local transport signatures and again suggest local incision and progradation.

To the north, oolitic ironstones occur at the base of a fossiliferous, wavy-bedded interval often containing brachiopod and echinoderm wackestones and packstones and relatively thin shales. These are interpreted to represent the SB. Southward into Iowa and eastward into northern Illinois, the same unit (the St. James sub-member of the Ion) shows less shale and becomes significantly more carbonate-dominated, and is heavily dolomitized in northern Illinois. The sudden appearance of carbonates at the base of the Cummingsville (and the St. James) and the upward-cleaning of facies represents the TST of the M5B sequence. Upward the St. James contains pyrite and apatite-impregnated dark hardground surfaces, many of which carry abundant assemblages of *Prasopora* bryozoans and other bryozoan taxa. An especially prominent contact and *Prasopora* zonule occurs just above a hardground near the base of the Beecher Member of the Dunleith Formation. Shales in this interval again begin to show enriched neodymium values characteristic of the Taconic Foreland and another carbon isotopic excursion (Fanton & Holmden, 2007) and thus the succession is interpreted as the latest TST to lowest HST interval of the M5B sequence. Above the prominent regional maximum flooding zone, distinct shales become somewhat less dominant and neodymium isotopic values again begin to drop. Shales are especially less predominant in the south where facies become more highly bioturbated and show evidence of abundant *Thalassinoides*-type burrow networks. The more extensive bioturbation is similar to patterns observed elsewhere as representative of the M5B HST.

Small rugose corals return to the faunal assemblages of the upper Beecher Member and the overlying Eagle Point Member as well as abundant receptaculitid algae. Thus suggests prominent shallowing and even possibly significant warming during the late M5B HST/RST. The top of the M5B sequence is not yet clearly established; however, the Eagle Point may represent the very latest HST/RST deposit and the shallowest facies between the two sequences.

Witzke and Bunker (1996) suggested that the upper contact of their TR cycle 5A was coincident with the top of the last shales of the Decorah just below the base of the Rivoli Member of the Dunleith Formation (interpreted to be base M6A sequence herein). Thus Decorah-style shale deposition ensued for at least one additional sequence (through the top of M5C) in the Hollandale Embayment area, although carbonate-dominated deposition continued to dominate in the south.

Cincinnati Arch

As discussed by Brett and colleagues (2004), the M5B sequence in the Cincinnati Arch/Jessamine Dome coincides with the lower to middle Grier Member of the Lexington Limestone. The Grier has been subdivided into three sub-units (see chapter 5) all below the level of the Macedonia Beds, the lowest two are considered the TST and HST of the M5B sequence respectively. In the shallowest water environments as recorded near Frankfort, Kentucky and in the vicinity of Herrington Lake, calcarenite facies and skeletal grainstones and rudstones are characteristic of the lower Grier TST and sit sharply above the upper Logana SB. These facies grade laterally into nodular wacke to packstones in slightly deeper water settings off the top of local uplifts. Weakly developed condensed beds have been noted as have been a number of hardground surfaces although they are not as prominent as in other units.

The middle sub-unit of the Grier is, however, a somewhat shalier facies, dominated by argillaceous burrow-nodular wackestones and packstones bearing a diverse fauna including *Prasopora* and large crinoids. This facies is very similar to the Sugar River of New York. Northward of Frankfort, the middle sub-unit is fairly uniform in lithology through the Cincinnati region before it loses its burrow-nodular appearance and transitions on the margins of the Sebree Trough into a rhythmically-bedded calcisiltite-shale facies similar to the Logana. Thus, this middle Grier is considered to represent the HST of the M5B sequence. It is generally a shallower HST facies than the underlying Logana, although northward the deeper facies is present. Overall the Grier is noted for its relatively high phosphate content that usually occurs as disseminated phosphate and/or as infillings in bryozoan zooecia, crinoid plates, and as coatings on shells. Most phosphatization occurs some distance above the base of the lower Grier and upward into the formation, although specific horizons of more enriched phosphatization are not, as yet, identified. This phosphatization suggests at least some upward condensation and sediment starvation in the M5B TST. The top of the M5B sequence is coincident with the change to the upper Grier sub-unit which again becomes somewhat more coarse-grained and calcarenitic. The base of the change is interpreted as the SB between the M5B and the M5C sequences.

Central Pennsylvania

The M5B sequence in central Pennsylvania is recognized here by the middle to upper portion of the New Enterprise Member of the Salona Formation. The lowest soft-sediment deformation intervals coincident with those in the Logana of Kentucky, the bentonite swarms in the basal Salona (K-bentonites 0 – 4), and the prominent graptolite-bearing beds immediately above K-bentonite 4 (Ashcom Shales) are included in the progradational phase of the HST of the underlying M5A sequence. Above this level, the New Enterprise Member shows a loss of thicker shale interbeds and a slight increase in grain size through an interval of between four and five meters. In this tripartite interval, calcisiltites become interbedded with fine-grained grainstones that show an increased occurrence of cross-stratification including tabular and HCS types, graded bedding, and slightly more fossiliferous beds. A thin shaly zone in the middle of the succession splits the unit into three parts. Facies in the upper and lower zones are nearly identical, however, occasional pyritic, organic-rich hardgrounds are developed on some bedding planes near the top of the unit. These grade upward rather sharply back into calcisiltite facies interbedded once again with thicker shale beds that contain yet another interval of soft-sediment deformation below the Roaring Spring Member.

Thus, although much of the New Enterprise is a "rhythmite" facies, the slight coarsening and loss of shale in the middle of the member is inferred to represent a base-level rise event and a weak TST. As is typical of most TST's, this TST shows evidence of a tripartite succession, with the uppermost beds showing evidence of increased condensation and mineralization during the late transgression. The ensuing shalier upper New Enterprise is interpreted to represent the HST of the M5B and renewed period of progradation and shale deposition, the latter of which is especially prominent in the subsurface of western Pennsylvania as discussed by Wagner (1966). Although dominant in much of the lower New Enterprise Member, *Cryptolithus* is absent in the uppermost New Enterprise suggesting shallowing of facies before deposition of the overlying Roaring Spring Member. The top of the M5B sequence is interpreted to occur at or near the base of the Roaring Spring Member where cross-bedded calcarenites or fine-grained grainstones and packstones become once again prevalent in the succession of interbedded calcisiltites and calcilutites (see below).

M5C Sequence: Rathbun–Poland; U. Grier-Macedonia; Roaring Spring

In the type region of New York State, the first significantly abrupt lateral facies changes are observed in the M5C sequence (**figure 16**), beginning with the Rathbun Member of the upper Sugar River Formation and continue upward into the overlying Denley Formation and higher sequences on the Trenton Shelf. This appears to be a pattern that is shared with Kentucky and

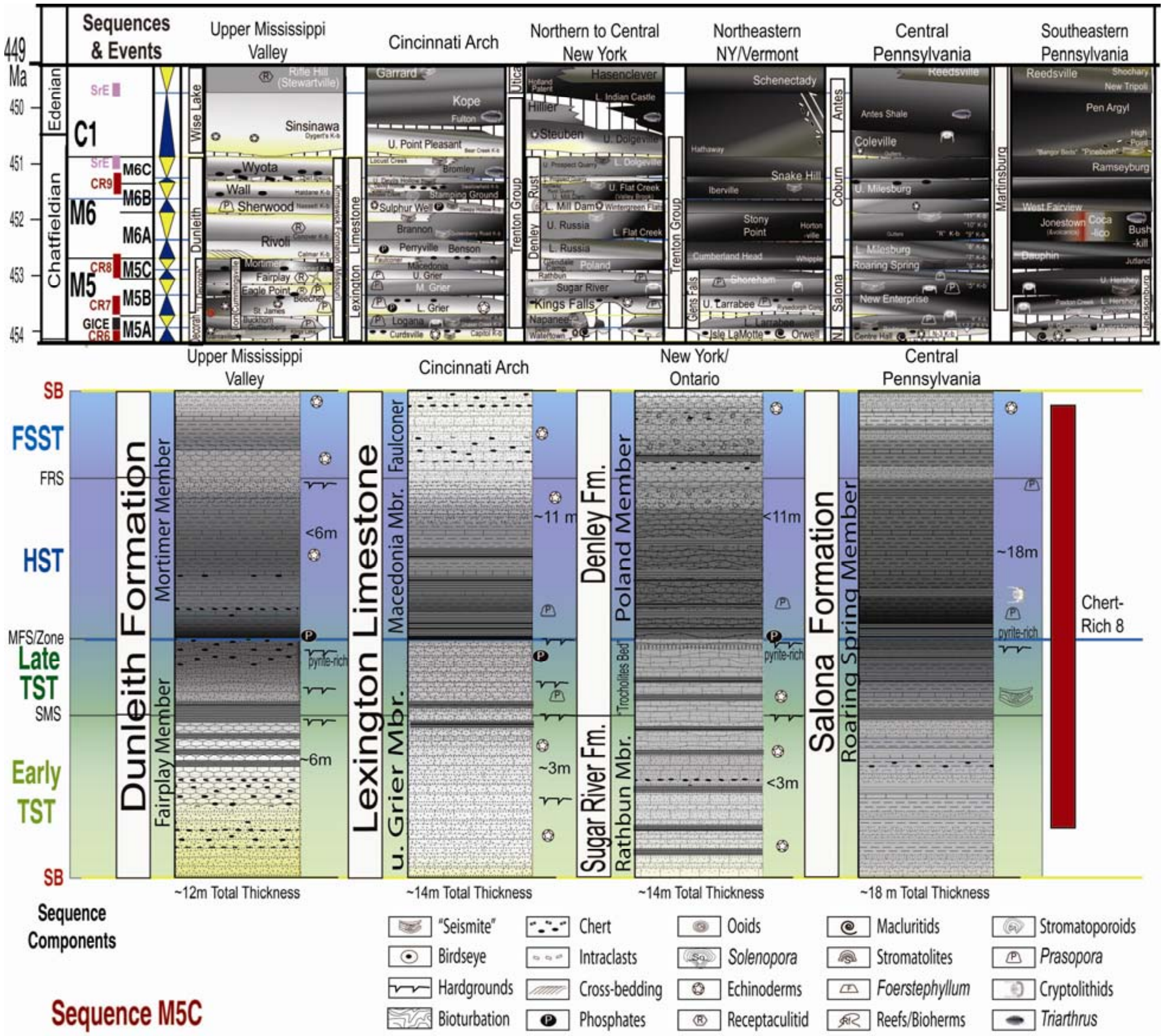


Figure 16: Correlated framework of the M5C sequence (Mid Chatfieldian sequence (Early Shermanian)) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

with central Pennsylvania. The basal SB of the M5C sequence is generally subtle in most regions of the northwestern Trenton Shelf. However in the Mohawk Valley as at Trenton Falls on West Canada Creek, the dramatic change to crinoidal grainstones stands in sharp contrast to the underlying shaly-nodular fine-grained carbonates of the lower Sugar River member. As these coarser grained carbonates are traced both east and west of Middleville, the facies grade into finer-grained carbonates that are difficult to separate from the remainder of the Sugar River. Hence, the development of the Rathbun Member grainstones is relatively localized to the

"Middleville Arch" and appears to be related to tectonic uplift of portions of the Trenton Shelf late in the M5B sequence. Here the base of these facies is interpreted as the SB. Upward, Rathbun grainstones and their lateral equivalents shows upward deepening patterns as recorded by increased evidence of condensation and sediment starvation upward into the late TST and to the maximum flooding zone.

The MFS interval of the M5C sequence is fairly well-established at the base of the Poland Member of the Denley Formation. A distinctive set of amalgamated condensed beds is well-developed (the Glendale sub-member), that continue the upward-deepening pattern of the Rathbun. These are capped by the substantially more condensed City Brook (*Trocholites)* bed that has an upper surface displaying phosphate and pyrite staining. This contact is thus interpreted as the MFS of this sequence. When correlated eastward in the central Mohawk Valley area, this contact passes abruptly, into dark gray to black calcareous, and graptolite-bearing shales of the Flat Creek Formation characteristic of the Taconic foredeep basin.

To the west, on the Trenton Platform, above the basal Glendale sub-member, the remainder of the Poland Member of the Denley Formation shows an upward coarsening series of small-scale cycles which demonstrate a progradational stacking pattern and a shallowing in faunal assemblages (see chapter 3). Within the upper part of the Poland a series of amalgamated packstone and fine-grained grainstones (although shaly) represent a rapid shallowing and are thus interpreted as the transition out of the lower HST and into the RST interval. The upper sequence boundary of the M5C sequence appears above the base of these amalgamated packstones and is coincident with the change into the lower Russia Member.

CORRELATION OF THE M5C SEQUENCE

Upper Mississippi Valley

The M5C sequence is here correlated roughly with the uppermost portion of TR cycle 5A as recognized by Witzke and Bunker (1996). This interval was described as containing three to four generalized pulses of shale progradation (regressive phases), each punctuated by phosphatic lag horizons and/or oolitic ironstone units (interpreted to be transgressive intervals). The earliest of these pulses correspond to the M5B sequence and the uppermost to the M5C sequence recognized here. Specific details of these depositional sequences are lacking in previous publications; however, the sequence coincides with the uppermost Cummingsville Member of the Galena Group of Minnesota or its equivalents including the uppermost Decorah of Iowa and portions of the middle Dunleith Formation of Wisconsin, Illinois and northeastern Iowa. In the latter regions, the Fairplay Member of the Dunleith forms the TST, and the overlying Mortimer Member forms the HST and represents the uppermost pulse of Decorah-type shale just below the widespread and relatively pure carbonates of the Rivoli Member of the Dunleith Formation. The latter of which is here interpreted as the M6A sequence TST (TR cycle 5B of Bunker and Witzke, 1996).

Witzke and Bunker (1996) define the Fairplay as a tripartite dolostone-dominated unit (in Iowa). Before dolomitization, the unit was evidently dominated by medium to thick-bedded, burrow-mottled wackestones (*Thalassinoides* are abundant) and packstones. Beds at the base and top of the unit contain coarser-grained stringers with abundant brachiopods and occasional cherts. The unit is also known to contain receptaculitid algae and horn corals (often silicified) that are thought to reflect relatively shallow-water environments. The unit becomes somewhat more argillaceous and recessively-weathering towards the top where beds are generally sparsely fossiliferous and more planar-bedded. Some beds, however, do show thin condensed brachiopod and crinoid-rich coquinal stringers. Occasional hardgrounds have been noted to occur within the Fairplay including near the top of the unit. Thus given this pattern, this unit is considered to represent the TST of the M5C sequence.

The HST of the M5C sequence in this region is represented by the Mortimer Member of the Dunleith Formation. The unit was sub-divided informally by Witzke and Ludvigson (2005) into lower, middle, and upper parts. Its base is an argillaceous mudstone to wackestone facies that is often wavy-bedded and interbedded with greenish gray shales (sub-rhythmite facies). Shales are most abundant in the lower portion of the lower sub-unit although they are interbedded with wackestones (fossiliferous calcisiltites) that often show small chert nodules. Fossils become more abundant upward and include scattered crinoid fragments, brachiopods and occasional gastropods. The middle sub-unit is dominated by irregularly-bedded, skeletal wackestones to packstones and reflects slightly shallower deposition. Cherts become more common and are occasionally bedded in nature. Near the top of the middle sub-unit packstones become more dominant and exhibit more extensive bioturbation and a sharp hardground contact (possible FRS or another minor flooding surface). The third sub-unit of the Mortimer Member is again more argillaceous and resembles the lower sub-unit, except that it is generally more fossiliferous and includes numerous echinoderm remains, scattered bryozoans, and sowerbyellid brachiopods. The Mortimer has been shown to contain exceptionally preserved echinoderm faunas in Iowa, Minnesota, and Illinois (Brower, 1992a,b, 1999).

Overall, the occurrence of significant shaly intervals of the Mortimer represents the final progradation of shales associated with HST to RST facies of the TR-Cycle5A sequence. In fact, neodymium isotopic values of this portion of the Dunleith (post I-3 pre-I-4 excursion of Fanton

& Holmden, 2007) show a return to more local provenance before once again becoming dominated by shales from the Taconic (in the TST of the M6A sequence). The top of the M5C sequence in this region is recognized at the distinctive base of the Rivoli Member.

Cincinnati Arch

In the Cincinnati Arch region, the M5C sequence is represented by the upper Grier (TST) and the Macedonia Beds and overlying Faulconer Member of the Lexington Limestone (early and Late HST). The uppermost Grier is composed of up to three meters of skeletal calcarenites and rudstone facies immediately below the Macedonia Beds. These coarse-grained facies show an upward change into more condensed facies and are interpreted as the TST of the M5C sequence. A particularly prominent zone near the top of the Grier and just below the base of the Macedonia is shown to have an abundance of nautiloids preserved on some bedding planes. Some nautiloids in this zone show evidence of reworking after early diagenesis and are typical of condensed zones and hardground producing intervals in other sequences.

The HST of the M5C coincides with the base of the Macedonia Beds. Like the Mortimer Member, the Macedonia can also be described as a tripartite succession. Sharply overlying the coarse skeletal grainstones and rudstones of the underlying Grier, the first homogeneous calcisiltites interbedded with calcareous shales appear above the MFS. The lower Macedonia beds grade upward into coarser-grained skeletal wackestone to rudstone (middle Macedonia beds) and then back into another succession of calcisiltites and shales (upper Macedonia beds). The Macedonia interval shows some cherts representative of chert-rich interval 8.

Overall, the Macedonia is another rhythmite facies characterized by tabular to lenticular-bedded argillaceous calcisiltites and interbedded shales with occasional fine-grained calcarenite stringers. The latter of which are more prevalent in the southern Jessamine Dome. As discussed in chapter 5, the Macedonia shallows upwards into beds containing hummocky cross-stratification and minor graded beds, especially in the southwestern Jessamine Dome where *Tetradium* and ostracods have been identified from the unit. Nonetheless, as with the M5C sequence in New York, lateral facies change is significant in this unit. Northward toward the Sebree Trough, the Macedonia is predominantly a shaly nodular facies that rapidly grades laterally into rhythmite facies in the central to northern Jessamine Dome. In the area of Cincinnati, the Macedonia loses its interbedded calcisilts and becomes a shale dominated facies with only minor carbonate interbeds and represents the most basinal facies of the region. The gradual shallowing shown to the north, and the prominent shallowing to high-energy facies of the Faulconer Member in the southern Jessamine Dome region indicates the latest HST/RST interval of the sequence. The Faulconer Bed RST is a rather massive, light gray, coarsely crystalline, fossiliferous limestone often containing highly abraded and fragmented fossil debris and includes occasional colonies of *Tetradium* corals. The association suggests rapid shallowing over the Macedonia and is typical of a RST facies. The SB at the top of the M5C is drawn at the base of the fine-grained calcilutite facies showing fenestral birdseye lithologies of the Salvisa Member deposited during the TST of the M6A sequence.

Central Pennsylvania

The M5C sequence in central Pennsylvania is represented by the minor shallowing of facies out of the lower Salona Formation. Slight shallowing is recorded during deposition of the Roaring Spring Member of the Salona Formation, which itself contains evidence of a TST and HST phase. Overall, the Roaring Spring is defined by the greater proportion of laminated-to-cross-laminated, fine-grained calcarenites interbedded with calcilutites and shales in the upper part of the Salona compared to the underlying New Enterprise. The calcarenites or fine-grained grainstones and packstones are similar to the TST facies of the M5B sequence, and are still relatively minor components of the interval. The coarse-grained facies represent only about 20% of the unit.

The base of the Roaring Spring Member, and the approximate base of the M5C sequence, is drawn at the level of the first abundant cross-bedded calcarenites, which first occur just below the level of the Salona 5 K-bentonite. The ensuing interval shows an abundance of highly repetitive cycles of thin- to medium-bedded calcilutite beds (occasionally peloidal) interbedded with fine-grained slightly fossiliferous grainstone beds. Cycles are typically separated by very thin calcareous shales. Upward these beds once again pick up *Cryptolithus* trilobites and demonstrate a number of condensed coquinal rudstone beds and at least one prominent pyrite-rich and iron-stained hardground contact that demarcate the change to the upper Roaring Spring Member. This interval is thus considered to represent the TST and MFS of the sequence.

The HST succession of the M5C sequence is coincident with the relative increase in thickness of shale interbeds and an associated thinning of calcilutite beds out of the lower Roaring Spring Member. When correlated laterally toward the southeast across the Adirondack Arch, the Roaring Spring becomes significantly more argillaceous and somewhat siltier where it transitions into Martinsburg lithologies in the Cumberland Valley. To the west into westernmost Pennsylvania, the Roaring Spring facies also grade into shale dominated facies similar to the Flat Creek shales of New York, and the distal Macedonia equivalent in the Sebree Trough of Kentucky. The appearance of shales over much of the region reflects a major pulse in the progradation of siliciclastics and may reflect initiation of a significant base-level change.

As in New York, the M5C HST shows a number of small-scale parasequences that generally form a tripartite succession similar to that observed elsewhere in this same sequence. The upper and lower succession are dominated by rather rhythmically bedded calcilutites, but the middle succession is slightly coarser-grained and demonstrates a number of more closely spaced calcarenite beds before once again becoming quite shaly below the Coburn Formation as demarcated by the sudden appearance of fossiliferous *Sowerbyella*-bearing wackestones, packstones, and brachiopod-rich rudstones interbedded with black shales and calcilutites. These fossiliferous beds stand in sharp contrast to the underlying Roaring Spring Member and are inferred to represent either the latest HST/RST or SB interval of the M6A sequence.

M6A Sequence: Russia-L. Rust; Perryville-Brannon: L. Milesburg

In New York State, Brett and colleagues (2004) recognized the transition out of the upper Poland Member into the Russia Member of the Denley Formation as a very condensed TST prior to the pronounced deepening event associated with the M6A sequence HST (**figure 17**). The deposition of very pure, calcilutite facies with very little accessory shale on the Trenton Shelf is interpreted to represent a pronounced period of siliciclastic sediment starvation during initiation of the TST. The capping beds of the Poland are distinctly non-fossiliferous and their cap is accentuated by the deposition of the twin Kuyahoora K-bentonites that appear to have been preserved on flooding surface contacts. Unlike the equivalent sequence in Kentucky, where a sharp karstic contact and SB are well-developed within the Perryville Member, the same SB in New York is very subtle as it is developed in substantially deeper water facies as is the case over much of the New York – Ontario region.

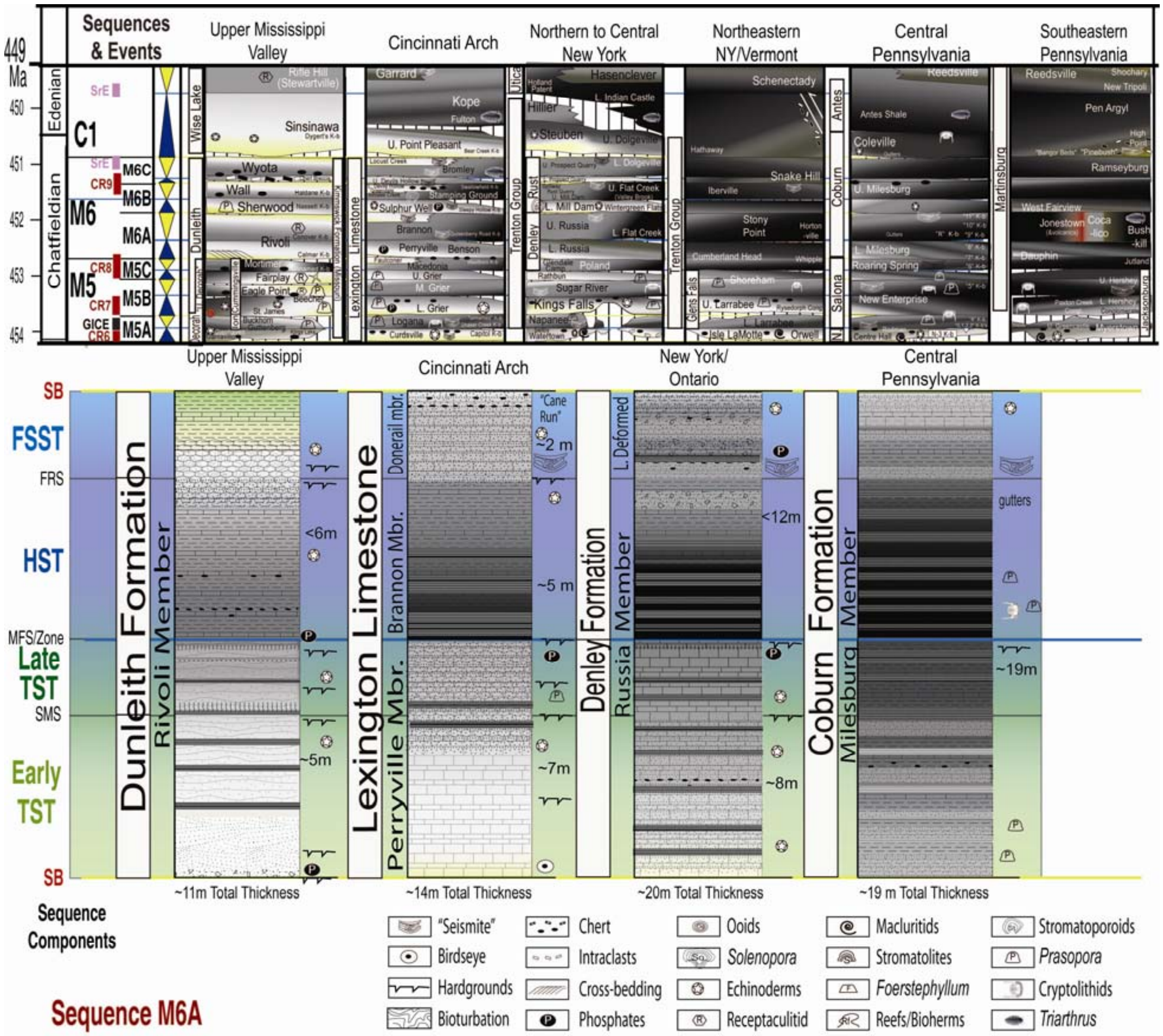


Figure 17: Correlated framework of the M6A sequence (Chatfieldian sequence (Mid Shermanian)) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

As highlighted by Brett and Baird (2002), the remainder of the lower Russia shows several upward-deepening, retrogradational cycles delineating the later portion of the TST. Moreover near the top of the lower Russia, evidence for condensation occurs again with the development of several amalgamated condensed beds and firm to hardground beds at the caps of several cycles ("overhanging ledge bed" and the "Castle Road bed"). Above the MFS, in New York, as in Kentucky (see below), the change into the M6A HST is well-defined by yet another succession of rhythmic-bedded calcilutites and shales of the middle to upper Russia Member.

The onlap of significant shale facies at this time in the middle to upper Russia records a significant deepening of the M6A HST and is recorded as a prominent, widespread surface recognized across the GACB.

The upper part of the M6A sequence is typically very well developed. The later part of the HST (going into the RST/FSST) is represented in the type Trenton Falls region, by the transition out of the Upper High Falls sub-member of the Russia, into the basal Rust Formation. The development of two unique cycles represented by the Taylor Fork Bed and the "Lower Disturbed Zone," signals a dramatic change to coarser grained limestones and significantly prograded shallow water depositional conditions in advance of the lower Mill Dam Member of the Rust Formation that records the M6B TST.

The occurrence of K-bentonites, widespread soft sediment deformation or "seismites" within the Upper High Falls sub-member of the Russia, and dramatically increased siliciclastic input on the Trenton Shelf has been interpreted to represent intensified tectonism during the later part of the M6A HST. Based on correlations of the upper Denley Formation downramp into siliciclastic dominated deposits, much of this tectonic development resulted in accentuation of steepened ramps, and rapid changes in lateral environmental gradients. Thus, it is not surprising that this interval also records a period of significant faunal change across much of the GACB as faunas are forced to relocate and/or go extinct.

CORRELATION OF THE M6A SEQUENCE

Upper Mississippi Valley

In the Upper Mississippi Valley, the TST of the M6A sequence corresponds with the initiation of TR Cycle 5B of Witzke and Bunker (1996). TR Cycle 5B correlates roughly with the M6 sequence recognized by Holland and Patzkowsky (1996, 1998) and represents a prolonged sequence from the base of the Middle Dunleith to the base of the Wise Lake Formation. Nonetheless, this sequence can be subdivided like the TR Cycle 5A below, into additional sequences that are comparable to those recognized by Brett and colleagues (2004).

The prominent change out of the siliciclastic-rich interval of TR Cycle 5A into carbonate-dominated facies (although still argillaceous) at the base of the Rivoli Member of the Dunleith Formation marks the onset of sequence M6A in this region. The expansion of somewhat cleaner carbonates in the middle of the Dunleith was interpreted, by Witzke and Bunker, as indicative of a renewed period of sea-level rise and flooding of siliciclastic producing source areas near the Transcontinental Arch. In Minnesota, the base of the Rivoli carbonate-dominated succession is marked by a prominent sandy phosphatic lag deposit representing the combined SB-TS and the lower TST interval. Some of the sands appear to be well-rounded and possibly frosted suggesting a possibility of aeolian transport. Toward the south into northern Iowa, the contact is recorded by a pair of phosphatic hardgrounds that mark the base of the transgressive limestones. In Iowa, the base of the Rivoli is typically a skeletal mudstone to wackestone unit with sculpted phosphatic stained hardgrounds. These facies grade upward into wackestones and packstones interbedded with argillaceous seams that become somewhat more pronounced upward. This interval represents the later TST/condensed interval and hardgrounds are more numerous and often include *Trypanites* borings and other faunas. Other fossils in the packstone beds include crinoids (in some places articulated) and a number of brachiopods, and gastropods that become more diminutive and less diverse upward. The succession of beds appears to be very condensed and contains at least one K-bentonite (the Calmar K-bentonite).

The HST of the M6A sequence in this region is represented by the upper Rivoli Member that can be divided into a tripartite succession (Witzke & Ludvigson, 2005). Overall, the lower and upper portions of the upper Rivoli show a greater proportion of green-gray shales and argillaceous skeletal wackestones. The middle unit is more massive-bedded and shows more pronounced hardgrounds likely representing the early-late HST boundary facies.

Although not as deep as the equivalent facies in the Mohawk Valley of New York, the lower middle-Rivoli shaly-interval represents the deepest facies of the sequence in this area. Neodymium isotopic values (event I4 of Fanton & Holmden, 2007) at about this level show a pronounced shift to Taconic signatures verifying the diminished influence of local clastic source areas. Bioturbation in the middle unit of the HST is more dominant than in the underlying TST although hardgrounds are still developed and characteristic– likely as a result of multiple smaller-scale depositional cycles that have not yet been formally recognized (but see Witzke and Ludvigson, 2005). Faunally, the HST shows abundant large trepostome bryozoans (potentially *Prasopora*), abundant *Isotelus* trilobites, small brachiopods, and *Chondrites* burrows. The uppermost Rivoli sub-unit shows a return to an interbedded argillaceous crinoid-bearing wackestones and packstones that also contain *Rafinesquina*, sowerbyellids, and various orthid brachiopods. Also again present are rugose corals and occasional receptaculitids that have been absent from the HST fauna. This faunal shift suggests a shallowing upward through the Rivoli. This is substantiated by the neodymium isotopic values that again drop back to values more characteristic of regional provenance.

The latest HST/RST interval is not yet clearly identified if it is present at all. However, the basal unit of the overlying Sherwood Member of the Dunleith shows a relatively massive, bioturbated skeletal wackestone dominated by crinoid debris and fragmented bryozoans and more numerous receptaculitid algae. It is unclear if this facies represents continued shallowing associated and therefore representative of the last portion of the M6A sequence, or if the facies represents the basal TST of the M6B sequence with a SB at its base. The overlying beds, however, do clearly show evidence once again for deepening. Thus here, the contact between the Rivoli and the overlying Sherwood are tentatively considered to represent the M6A-M6B sequence boundary interval.

Cincinnati Arch

In the Cincinnati Arch region, the M6A sequence is characterized by the first return to peritidal facies since the end of the Turinian (uppermost High Bridge Group in sequence M4), although the areal extent of the peritidal facies was restricted to the southwestern Jessamine Dome area. With steeply dipping ramps developed during the M5B and M5C sequences, facies within the M6A sequence continue to show rapid lateral facies change from the shallowest proximal facies on the Jessamine Dome into significantly deeper water distal facies in the Sebree Trough to the north. McLaughlin and Brett (2007) and Brett and colleagues (2004), suggested the M6A sequence TST was represented by the lower Salvisa Member of the Lexington Limestone. The base of the Salvisa has long been recognized as a sharp karstic contact separating the nodular facies of the Faulconer Member from the peritidal fenestral micrite facies of Salvisa. This contact was the same major sequence boundary that Holland and Patzkowsky (1996, 1998) used to designate the base of the M6 sequence. Farther north along the axis of the Cincinnati Arch, into the northern Jessamine Dome the same SB is not as pronounced and is more subtle in its expression due to the loss of the peritidal facies above it. Facies representing the M6A sequence TST in these more northerly areas (Benson Member of older workers) are more similar to the underlying Grier and are commonly included in that facies by other workers.

Nonetheless, as in New York the maximum flooding zone and HST of M6A sequence are represented by the thin-bedded, calcilutite and shale rhythmite facies of the Brannon Limestone Member. The pronounced change, even in the shallowest areas of the Jessamine Dome, records a major deepening analogous in scale to that seen in the Logana–Napanee of sequence M5A. As with the M5A sequence, the occurrence of K-bentonites coincident with widespread soft sediment deformation structures (seismites) within the HST of the M6A sequence signals a period of intensified Taconic tectonism across the GACB. A major difference, however, between the M5A and the M6A sequence is that depositional gradients evidently became substantially steeper and more pronounced in the M6A sequence. Narrowing of facies belts and an expansion of shale-dominated facies is characteristic both of the Sebree Trough region and the Taconic Foreland in New York.

The top of the M6A sequence in the Cincinnati Arch region is demarcated by a RST or FSST as represented by the "Donerail member" (McLaughlin et al., 2007) that often forms a soft-sediment deformed, channel-filling succession capped by a phosphatic bed characterizing the SB. As in New York where the Taylor Mill phosphatic bed erosionally overlies the channel filling succession of the Upper High Falls sub-member of the Russia Formation, the ensuing TST is again characterized by relatively pure and substantially coarser-grained carbonates reflecting the renewed transgression.

Central Pennsylvania

In the Ridge and Valley of central Pennsylvania, the M6A sequence is characterized by the prominent shift out of the Salona Formation into the lower Milesburg Member of the Coburn Formation. As in New York, the Upper Mississippi Valley, and the Cincinnati Arch, the M6A sequence contains an abundance of K-bentonites and abundant, deep-water shale facies. In central Pennsylvania, all facies are relatively deepwater facies and sequences components are not as prominent as in shallower shelf areas at this time. Nonetheless, sequence architecture is tentatively elucidated based on recognition of more subtle changes in lithology and faunal associations. Overall, as discussed in chapter 4, the base of the Coburn is drawn at the base of the first cross-bedded grainstones dominated by *Sowerbyella* and trilobite coquinas that are most prominent in the northern Ridge and Valley area (closest to the Trenton Shelf). The Milesburg coquinal beds appear as a wedge originating in the north and thinning laterally to the south and east across the Adirondack Arch into the Taconic Foreland. This pattern suggests that the Milesburg Member behaves much like a progradational unit originating in the north and prograding to the south and east and likely represents more than one third-order depositional sequence (here interpreted to represent the M6A and M6B sequences).

In sections where the Milesburg is best-developed, the lowermost beds below the R K-bentonite are relatively thin, cross-bedded calcarenites and *Sowerbyella* rudstones interbedded with thin shales. In some cases, shales partings are nonexistent where fine-grained calcarenites appear to be amalgamated by storm processes and winnowing. In many localities, the base of this interval contains bryozoan conglomerate deposits showing significant evidence for reworking and reorientation. Several meters above the top of these beds, and immediately below the R K-bentonite at Union Furnace, the unit once again becomes much shalier and contains abundant *Cryptolithus* trilobites, which in some cases are complete and show evidence of rapid burial. This succession is strongly suggestive of a HST facies with a relatively thin TST interval in the preceding interval. Above this level, cryptolithids and *Prasopora*-like bryozoans persist for a short interval before they are replaced upward by a sharp increase once again in the number and thickness of pack-to grainstone beds carrying a much more diverse fragmented *Sowerbyella*-

trilobite association recording the progradation of shallower facies during the latest HST to SB interval of the M6B sequence.

M6B Sequence: Mill Dam-Spillway Members; Sulphur Well-Stamping Ground; U. Milesburg

The M6B SB in the type region is a fairly sharp contact at the base of the Mill Dam Member of the Rust Formation at the upper contact of the "Lower Disturbed Zone" (**figure 18**).

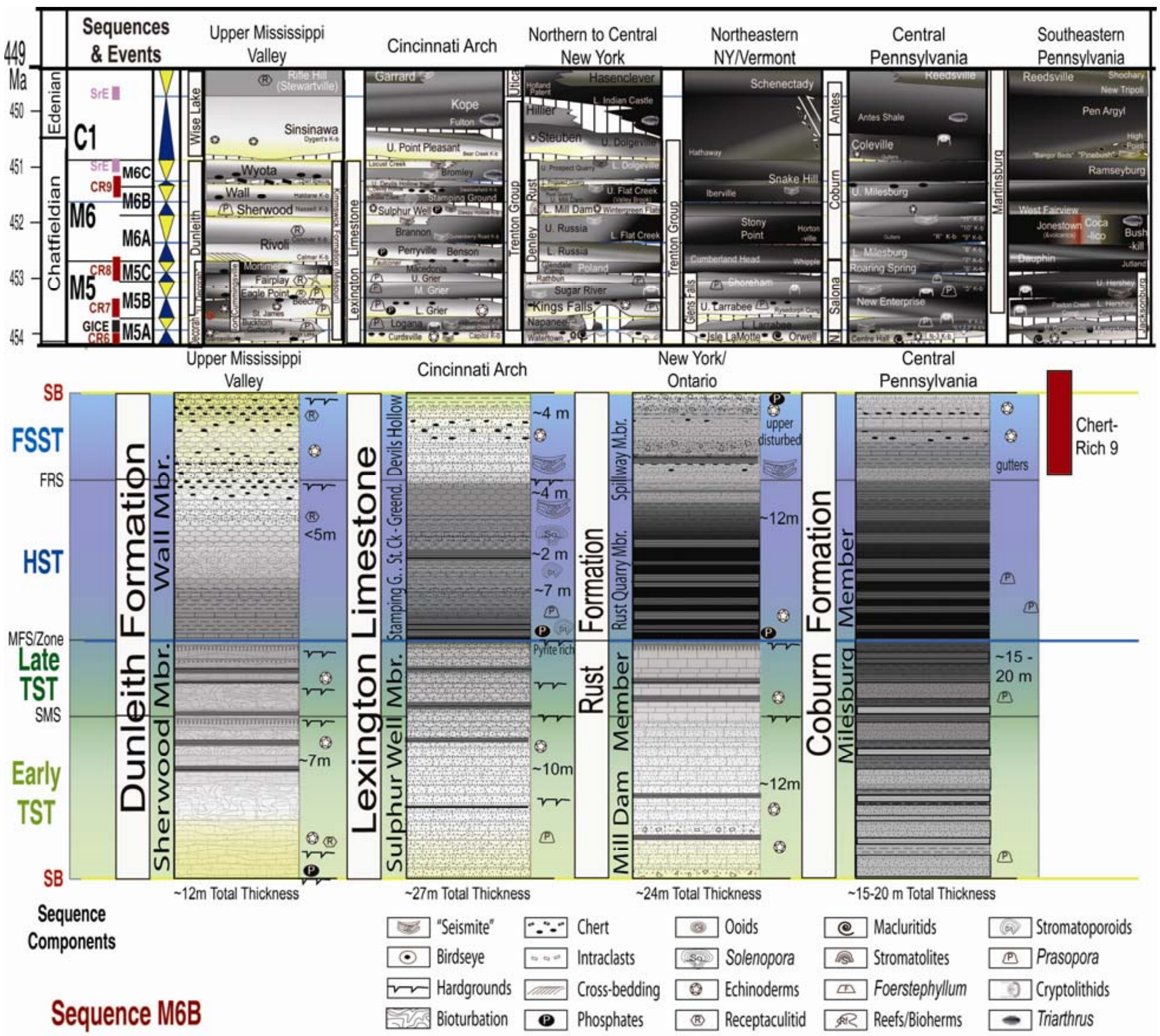


Figure 18: Correlated framework of the M6B sequence (Chatfieldian (Late Shermanian)) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

At this level, shaly nodular to brecciated fabrics of the disturbed zone transition to more massively-bedded coarse-grained packstones and grainstones of the Upper Mill Dam Member that represent the TST facies of the sequence. Within the Mill Dam the coarse-grained brachiopod and echinoderm facies show only minor argillaceous interbeds and cycles are poorly constrained. However, large *Rafinesquina* brachiopods as well as other fossil fragments show an upward change from abraded and fragmented textures to substantially more complete and pristine fossil specimens. This change in taphonomic signature, along with an increase in development of mineralized hardgrounds on small-scale cycle caps suggests that this interval is deepening and condensing-upward. The upper surface of the Mill Dam at the contact with the overlying Spillway Member in some localities in the Mohawk Valley is marked by pyrite and phosphate impregnated hardground showing evidence for corrosion. The surface is interpreted as an MFS.

Overlying the MFS of the M6B sequence, the Spillway Member of the Rust Formation is a shallowing-upward succession composed of several smaller-scale cycles, the lowest of which are composed of calcisiltites and shales. The lowest of these beds (referred to as the Rust Quarry member) often contain well-preserved trilobites previously studied by Wolcott. These facies show evidence for episodic obrution-style deposition and lagerstätte typical of HST intervals. These smaller-scale cycles become amalgamated upwards and gradually lose the thickest shaly interbeds, and across much of the shallow-water areas of the GACB shales are characteristically less abundant. As is common in most HST deposits, these cycles show a pronounced progradational stacking pattern. In this case the uppermost part of the Spillway Member (late HST) shows very dramatic evidence for channelization and slumping associated with rapid sea-level fall and a thin, but pronounced RST/FSST. The "upper disturbed zone" is located within the succession of the RST at Trenton Falls, and exhibits dramatically folded strata characteristic of “seismites” elsewhere. The unit is topped by a slightly phosphatic lag deposit (SB) and coarse-grained facies at the base of the overlying Prospect Quarry Member, here interpreted as the TST of the M6C sequence.

CORRELATION OF THE M6B SEQUENCE

Upper Mississippi Valley

In this region, the M6B sequence is recognized within the succession of TR Cycle 5B of Witzke and Bunker (1996). The M6B sequence corresponds with the deepening to regressive succession recorded in the Sherwood to Wall Members of the Dunleith Formation. It is suggested here that the Sherwood represents the TST and the Wall the HST facies, although much of the succession, including the overlying TR Cycle 5C (equivalent to M6C sequence herein) is becoming significantly starved of siliciclastics as is recognized by a weak neodymium isotopic event (I5 of Fanton & Holmden, 2007) at this time.

Overall Witzke and Ludvigson (2005) report the Sherwood Member to be an irregularly-bedded, burrowed skeletal wackestone with thin packstone lenses especially in the lowermost beds. Upward the unit becomes more fine-grained and argillaceous, and shows evidence for recessive-weathering notches and mineralized hardgrounds. Fossils are abundant in the unit and include crinoids, brachiopods (e.g. *Sowerbyella*, *Strophomena*), gastropods, large trepostome bryozoans, lingulids, abundant receptaculitids (only in the base), and fragments of calymenid trilobites. Generally faunas become somewhat less robust upward, but in some cases preservation is better in darker-colored facies (with increased TOC values as per Fanton & Holmden, 2007). Articulated crinoid stems are also observed in the unit despite some evidence for minor small-scale burrowing. Overall this succession suggests facies of the TST to maximum flooding zone.

The HST facies of this sequence is recorded by the Wall Member of the Dunleith. The Wall was described by Witzke and Ludvigson (2005) as dominated by skeletal mudstones to wackestones with minor mega-rippled packstones dominated by gastropods. The Wall shows abundant *Thalassinoides* burrows especially in the upper part and overall shows a coarsening-upward pattern with packstones and grainstones in lenses and thin-to medium beds becoming more abundant toward the top of the unit. Bedded and nodular cherts are also abundant in this uppermost facies and silicification of crinoid fragments is apparent and is coincident with the onset of chert-rich interval 9 that extends upward into the overlying M6C sequence. Receptaculitids and large gastropods become abundant at the top of the succession and at the very top of the Wall is a sharp contact interpreted as a hardground by Witzke and Ludvigson. This contact shows evidence of minor erosion and large burrow prods that are filled with packstone to grainstone materials. It is overlain by the basal Wyota Member of the Dunleith. Both lithologic and faunal evidence suggests a shallowing-upward succession with progradation of shallower, coarser-grained carbonate facies over finer-grained deeper water carbonates. This supports the interpretation of this interval as an early to late HST facies. Although more study is necessary, the prominent contact at the top of the succession is interpreted as the SB interval. However, the contact could represent a surface within the FSST to LST before the onset of the M6C TST. This may be substantiated by the slight increase in shale interbeds in the base of the Wyota Member that could represent the rapidly prograded FSST of the M6B sequence.

Cincinnati Arch

In Kentucky and Ohio, the M6B sequence is recorded by the prominent succession of coarse-grained facies often referred to as the Tanglewood Member of the Lexington Limestone. McLaughlin and Brett (2006) and Brett and colleagues (2004) recognize the succession to include the Sulphur Well, Stamping Ground, Strodes Creek, Greendale, and lower Devils Hollow Members of the Lexington Limestone. As in New York State, and the Upper Mississippi Valley the sequence continues to represent a shallower facies than much of the M5 composite sequence, although it does not shallow back to peritidal facies as did the underlying M6A sequence.

The SB of sequence M6B is a very prominent surface at the base of the Sulphur Well Member grainstones; there is evidence for erosive truncation of portions of the Donerail member beneath this contact. In the most proximal areas of the Jessamine Dome-Sebree Trough transect, the Sulphur Well is a well-bedded calcarenite facies (with numerous bryozoans). It grades laterally to the north and vertically into poorly-sorted bryozoan rudstone facies with minor calcisiltites in irregular to lenticular beds that are separated by thin shale partings. Northward, the Sulphur Well interval shows evidence of condensation and thins into Ohio where it can be traced into the shale-dominated succession of the Sebree Trough. The unit shows a back-stepping or retrogradational pattern and becomes deeper upward typical of a TST.

In proximal outcrop areas, the MFS interval is prominent, sharp and relatively planar with the coarse-grained calcarenites of the Sulphur Well overlain by the shaly nodular wackestones to packstones of the Stamping Ground Member. It is marked in most localities by pyrite and phosphate impregnated hardgrounds. This contact can be traced northward and eastward. In the latter region the contact is more subtle as the unit grades from bryozoan-rich limestone facies into bryozoan-rich shale facies in the base of the Millersburg Member (base Clays Ferry Formation of older workers).

The HST succession of the M6B sequence is recorded by the Stamping Ground – Strodes Creek – Greendale succession. Well known for its stromatoporoid biostromes, this interval clearly shows evidence for increased siliciclastics and an overall shallowing-upward, progradational pattern. In proximal areas, the Stamping Ground is a wavy nodular- to cross-bedded argillaceous packstone to grainstone unit containing abundant, overturned and fragmented stromatoporoids and occasional *Solenopora*. In many cases, the bryozoan-algal bioherms are draped by shales and argillaceous limestones in proximal areas.

Northward to the Ohio region, the HST passes out of nodular facies and into tabular rhythmite facies south of the margin of the Sebree Trough. In this region, the facies is representative of the typical Chatfieldian HST rhythmites. The later HST is represented by the Strodes Creek and its equivalents. Above the wavy nodular Stamping Ground, capping calcarenites are developed and contain a diverse but fragmented assemblage. Upward these sharply transition, across another rusty hardground contact back into more argillaceous facies of the Greendale Member. This succession is thus interpreted as the latest HST pre-RST/FSST strata (see McLaughlin et al., 2004). The Greendale is a dark shaly packstone and calcisiltite interval that overlies a distinct hardground that is heavily stained and mineralized similar to that developed at the top of the Sulphur Well. The contact is interpreted as a minor higher-order flooding event leading into the latest HST. The upper contact of the Greendale is unconformable and shows some local evidence for erosional truncation and channeling at the base of the lower Devils Hollow although it is typically flat and sharp. As suggested by McLaughlin and colleagues, this contact is likely a FRS with the overlying seismically deformed lower Devils

Hollow representing the RST/FSST. The uppermost M6B sequence FSST is thus represented by the lower Devils Hollow Member. The latter is a coarsely crystalline massive limestone containing abundant gastropod shells similar to the succession observed in the upper Wall of the Upper Mississippi Valley. The unit also contains a minor mud-cracked green shale in some localities. The occurrence of this mud, this late in the sequence may correlate with the muddy interval in the base of the Wyota Member of the Dunleith. Moreover, this facies is more characteristic of late HST/FSST siliciclastic-rich intervals of pre-Chatfieldian interval suggesting possible climatic signature with a possible wind-blown origin and/or increased precipitation and rates of runoff. The top of the M6B sequence is marked at the base of the Upper Devils Hollow where facies show a change to relatively clean micritic carbonates showing initiation of the next base-level rise.

Central Pennsylvania

The M6B sequence in central Pennsylvania is likely represented in the facies of the Upper Milesburg Member as mentioned above. Above the R K-bentonite, Upper Milesburg facies show a rapid progradation over underlying finer-grained units (uppermost Roaring Spring facies of Thompson, 1963). In contrast to M6A sequence succession, the M6B sequence above the R K-bentonite is characteristically coarser-grained and has more robust packstone to grainstone beds. The interval is typically thicker-bedded than the lower Milesburg and has a much more uniform thickness overall. In southern and central Valley and Ridge outcrops, the lower Milesburg is not present as a coarse-grained facies and only the upper Milesburg is developed above facies assigned to the Roaring Spring Member of the Salona Formation (although these beds are likely deeper water equivalents of the lower Milesburg). Thus, the succession appears to contain a weakly developed TST to HST succession. Although internal surfaces and systems tracts are not easily recognized, a general upward increase in dark shale partings is noted. The shale increase continues upward into the next sequence and is likely related to renewed tectonic subsidence in this region as the foredeep basin moved into the eastern Ridge and Valley region just to the east. The Milesburg does have a chert-rich interval (#9) at the top of the succession near the base of the Coleville Member and a prominent change to crinoid-dominated assemblages thought to reflect subsequent sequences. Thus, this chert likely occurs near the base of the M6C sequence although the contact interval is in need of additional study.

M6C Sequence: Rust Formation (Prospect Quarry Member)

The M6C succession in the Trenton Falls region of New York begins with a very thin zone of grainstones at the base of the Prospect Quarry Member where they sharply overlie and infill structural depressions formed on top of the Spillway deformed interval (**figure 19**). Both the top and bottom contacts of the Spillway deformed interval represent depositional discontinuities, which in the case of the upper contact is interpreted as the M6B-M6C SB.

The thin grainstone interval at the base of the Prospect Quarry Member shows dramatic evidence for upward condensation, and is capped by a pebbly phosphatic intraclast bed that is sharply overlain by thin-bedded, shaly calcilutite and fine-grained grainstones of the upper Prospect Quarry. This thin coarse-grained limestone package to shaly interbedded interval is interpreted as the TST, MFS, and early HST of the M6C sequence. The HST interval of this sequence is fairly thin (compared to Kentucky) but shows, as with lower sequences, an upward shallowing pattern out of the fine-grained shaly lutite cycles into bioturbated nodular wackestone

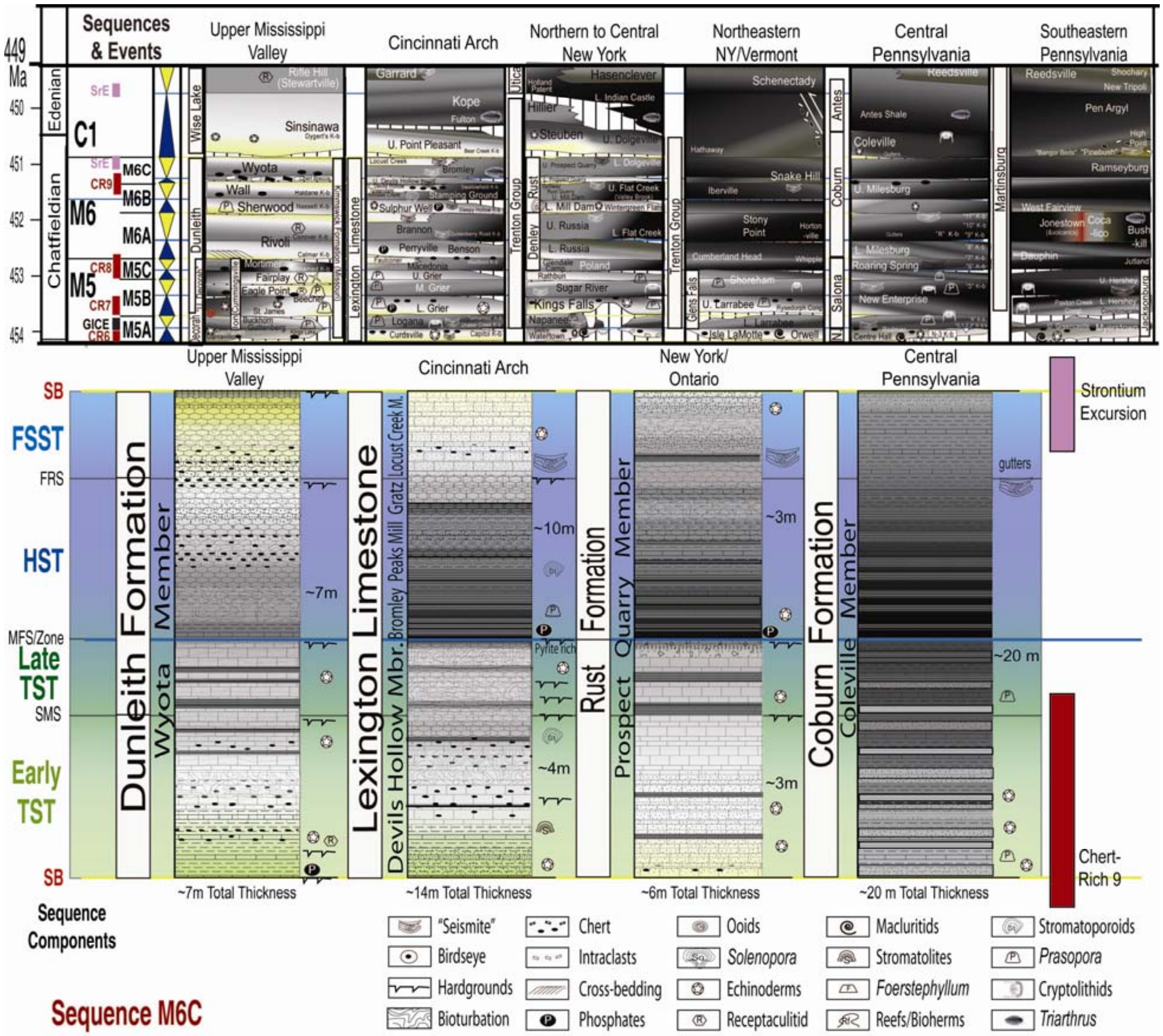


Figure 19: Correlated framework of the M6C sequence (last Chatfieldian sequence (Late Shermanian)) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

to packstone facies of the upper Prospect Quarry Member. The top of the Prospect Quarry Member of the Rust Formation and its lateral equivalent in the lower Dolgeville Formation is marked by the change upward into regressive fine-grained grainstones that rapidly prograde into the Taconic foreland. These facies are interpreted as RST/FSST facies in similar fashion to the underlying M6B sequence. And like the underlying sequence, this interval also contains significant evidence for soft-sediment "seismite" deformation in both regions immediately prior to deposition of coarse-grained facies of the Steuben Limestone and its equivalents.

CORRELATION OF THE M6C SEQUENCE

Upper Mississippi Valley

As with the M6B sequence, the M6C sequence in the Upper Mississippi Valley is represented within TR Cycle 5B of Witzke and Bunker (1996). It is the last depositional episode immediately prior to TR Cycle 5C recognized by these authors to begin at the base of the Wise Lake Formation. Interpreted here, the M6C sequence is thought to be represented by the strata contained within the Wyota Member of the Dunleith Formation. As mentioned above, it may be possible that the argillaceous base of the Wyota represents the FSST interval of the M6B sequence and may be an equivalent of the Greendale of Kentucky. However above this interval the Wyota again shows evidence of multiple high-order depositional cycles showing deepening and shallowing.

Using the descriptions of Witzke and Ludvigson (2005), the TST interval of the Wyota is thought to be represented by the change upward from fine-grained grainstones irregularly-bedded with skeletal wackestones into facies interbedded with lenses and thin beds of packstones containing abundant cherts, with crinoids, and rugose corals (among other fossils). Most beds in this unit are relatively thin to medium-bedded, but upward show more pronounced recessive-weathering, slightly argillaceous notches. Packstone beds also become darker in this interval and appear to be much more condensed than underlying beds. The interval also contains laminated fine-grained grainstones and stacked condensed packstones with a sculpted hardground surface at the top of the succession– here interpreted as the maximum flooding zone.

The upper Wyota in turn is thought to represent the HST of the sequence. A sharp change from packstone-bearing facies back into bioturbated, argillaceous skeletal lime mudstone and wackestone facies is representative of this maximum deepening and an overall increase in siliciclastic components (although still slight). Cherts are still highly abundant in this interval and in fact make up a significant proportion of the unit, often 40 to 50 %. This finer facies once again coarsens to the top of the Dunleith where rippled packstones and fine- to medium-grained packstones become abundant once again and large brachiopods become abundant. Although these facies continue upward into the Sinsinawa Member of the Wise Lake Formation, the uppermost Wyota beds are capped by a prominent darkened hardground surface with deep penetrating vertical burrows and burrow prods developed at the contact. This contact is tentatively assigned to the M6C-C1 sequence boundary interval although clearly more work is needed.

Cincinnati Arch

The uppermost Chatfieldian sequence in the Cincinnati Arch region is represented by the upper Devils Hollow, Bromley, Peaks Mill, Gratz, and Locust Creek Members of the Lexington Limestone/ Point Pleasant Formation (of Ohio). The TST of the M6C sequence is represented by the upper Devils Hollow, which is distinctly finer-grained than the underlying lower Devils Hollow. These upper beds show excellent cyclic bedding of calcilutites-lime mudstone facies – somewhat similar to the lower Wyota. They also contain ostracods and gastropods and possible cyanobacterial mats (evidenced by crinkly laminations on some bedding planes) suggesting a somewhat restricted association. Burrow mottling becomes more pronounced upward in the TST and beds are often slightly coarser, carrying fragments of crinoids, bryozoans, and other fossils. The upper contact of the Devils Hollow is recognized by the close stacking of as many as four different hardground surfaces that show evidence for sediment starvation and mineralization in the late TST condensed interval.

The uppermost hardground (MFS) is sharply overlain by the Bromley Shale and its characteristic dark gray shaly rhythmite facies, although laterally this unit shows significant gradation as discussed in chapter 5. In the Cincinnati region, the Bromley is typically a calcisiltite and shale rhythmite facies containing the first *Triarthrus* trilobites. The Bromley becomes slightly more nodular and interbedded with coarser calcarenite beds to the south into the northern Jessamine Dome. Northward into the Sebree Trough, the unit is a shale-dominated facies representative of the progradational nature of the M6C sequence where it has been called the "Utica Shale." As in the upper Wyota, the coarser-grained beds of the Bromley in the Jessamine Dome region often contain large brachiopods (*Rafinesquina*) although only minor evidence for chert development is observed.

Several stacked, small-scale cycles show development of recognizable hardgrounds and minor flooding surfaces in the Peaks Mill and Gratz Members. However, these gradually shallow-upward and the succession represents the Bromley HST. The uppermost facies show evidence of another sharp and often channeled erosion surface between the deep-water rhythmite facies and the shallow water calcarenite facies of the Locust Creek Member (McLaughlin and Brett, 2006). As with underlying sequences, this RST/FSST interval, as it has been interpreted, is yet again recognized to contain abundant deformed argillaceous calcarenites that are capped by massive coarse-grainstones of the basal Point Pleasant representing the C1 TST.

Central Pennsylvania

The M6C sequence in central Pennsylvania is here recognized to coincide with the rapidly deepening facies of the Coleville Member of the Coburn Formation. This interval is very similar overall to the Bromley Shale of Kentucky/Ohio and the Dolgeville Formation of New

York with which it is correlated in part. Characteristic thin-bedded fine-grained calcarenites, fossiliferous packstones, and calcilutites are increasingly interbedded with organic-rich, calcareous shales. The latter of which become more dominant to the south and east of the type region in the northern exposures of the Ridge and Valley.

Overall the base of the Coleville (shaly fine-grained calcarenites and packstones) is the most widespread portion of the member and likely represents the LST to TST facies of the M6C sequence in this region. The unit subsequently retrogrades successively upward into the C1 sequence where maximum flooding (of the C1) coincides with the first incursion of *Triarthrus* trilobites (which have yet to be found in the type Coburn). Further details are limited due to poor exposures and significantly faulted/folded successions within this shale-rich facies. However, in some localities large gutters are developed above an interval of possible syn-sedimentary deformed strata that can be used to divide the Coleville into upper and lower sub-units. This gutter producing interval may coincide with the M6C – C1 FSST to SB interval with the gutters representing the down slope equivalent of the FRS or SB. The upper Coleville is thus interpreted as the C1 TST (see below).

<u>C1 Sequence: Steuben to Hillier Formations, Upper Coleville, Pt. Pleasant</u>

Although pronounced changes in depositional sequence architecture occurred within strata deposited in central Pennsylvania in deepwater environments during the M6 sequences and arguably before, the C1 sequence shows the last stand of the GACB and a pronounced change in sequence architecture in shallow shelf regions for the first time (**figure 20**). Dramatic changes in sedimentation and basin configuration appear to have occurred during deposition of the C1 sequence in New York and elsewhere across the GACB. In the Trenton Falls to northern New

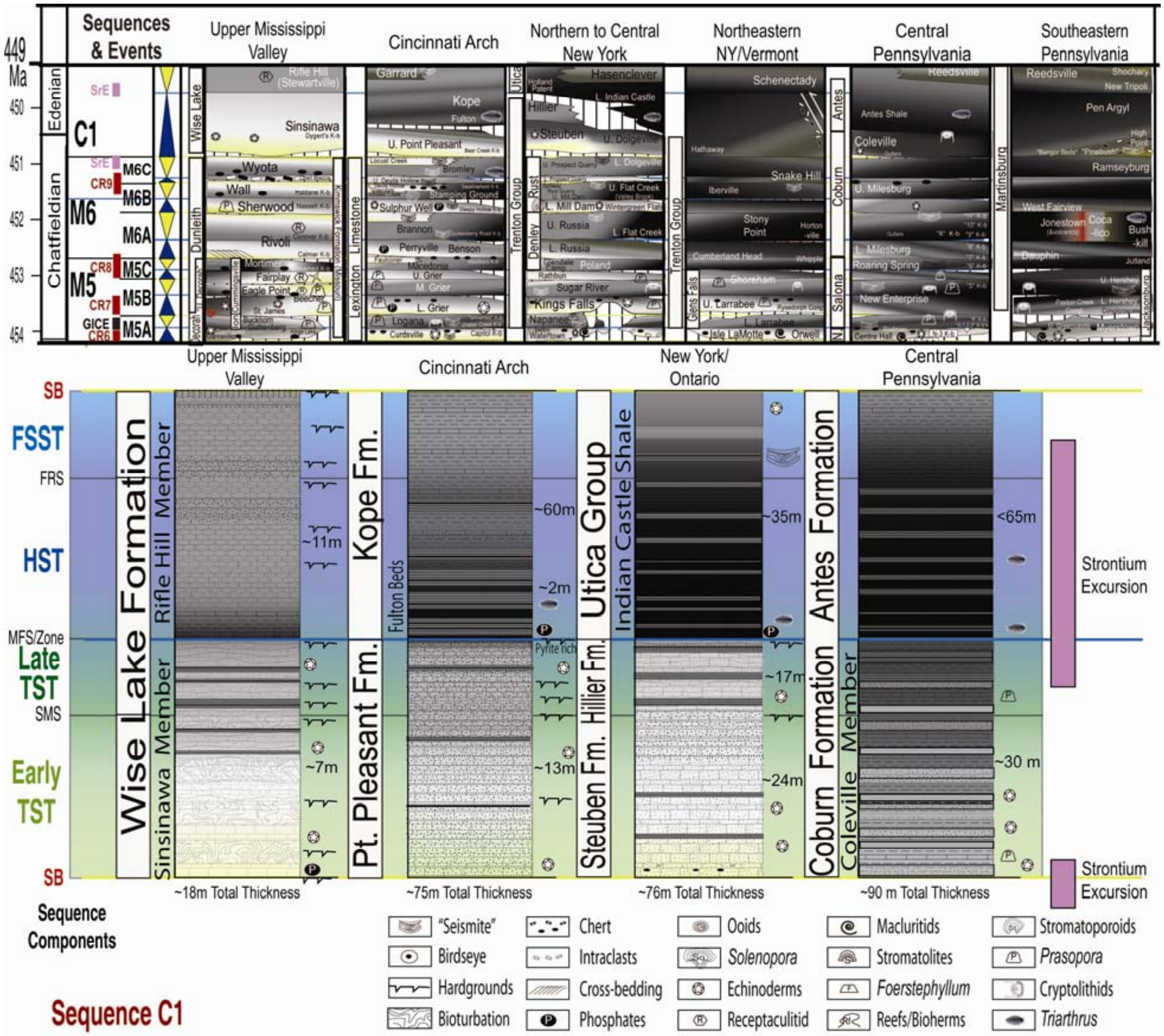


Figure 20: Correlated framework of the C1 sequence (Chatfieldian-Edenian Boundary Sequence) for the Upper Mississippi Valley, Cincinnati Arch, northeastern New York, and central Pennsylvania.

York region a widespread interval of crinoidal grainstone, referred to as the Steuben Limestone signals relatively shallow, but transgressive conditions, and sharply overlies the Rust Formation. With the development of cross-stratified crinoidal sand shoals, especially in the Trenton Falls region, the Steuben Formation most likely represents the TST component of the C1 sequence.

The uppermost Steuben passes both upward and basinward into a succession of fine-grained turbiditic calcarenites-calcisiltites and interbedded black shales, a facies that is very similar to the Coleville Member of the Coburn Formation of central Pennsylvania. The upper contact of the Steuben, shows an abrupt transition to a back-stepping succession of shales and argillaceous packstones and then into the major shale-dominated succession of the Indian Castle Shale (see Baird and Brett, 2002). The change out of the Steuben into turbiditic facies represents the late TST condensed interval and a lateral equivalent of the Hillier Formation of northern New York. This interval is so condensed within the Taconic Foredeep in the Mohawk Valley east of Trenton Falls that the dark shales of the Indian Castle rest on a major discontinuity marked by significant submarine dissolution and corrosion. Thus, the contact records a significant amount of time and is equivalent in duration to the late TST represented by deposition of the upper Steuben to Hillier Formations in more northwesterly areas. This MFS contact can be correlated upward onto the New York Shelf where the gap is significantly less, although shale deposition becomes pronounced during the ensuing HST. As the Steuben represents the culmination of the Trenton Group, further discussion of the sequence above is deferred (but see McLaughlin and Brett, 2007).

CORRELATION OF THE C1 SEQUENCE

Upper Mississippi Valley

In the Upper Mississippi Valley, the initiation of the C1 sequence is represented by the change from the Dunleith Formation into the Wise Lake Formation of Iowa and Wisconsin. This prominent contact was recognized by Witzke and Bunker (1996) to separate TR Cycle 5B from their TR Cycle 5C at the onset of the Edenian. TR Cycle 5C is equivalent to sequence G5 of Witzke and Ludvigson (2005). In Minnesota this contact is placed at the base of the Stewartville Formation (Sinsinawa Member). Although the lower Sinsinawa Member is characterized by wackestone facies similar to those seen in the lower Dunleith sequences (M5 sequences), Witzke and Bunker indicated that cycle 5C showed a transition to the purest carbonate lithologies of the

Galena Group. They suggested the significant loss of local siliciclastic source areas on the Transcontinental Arch likely resulted from a significant deepening pulse at this time. This siliciclastic-poor, wackestone-packstone facies, dominated by *Thalassinoides* burrow networks, generally continues upward in this region into the Dubuque Formation, although several pronounced shale depositional episodes do punctuate this pattern. As indicated by these latter authors, the basal Wise Lake sub-interval contains numerous hardgrounds, and generally is thought to represent the deepest facies of the entire cycle and is followed by small-scale cycles that gradually shallow upward. The basal six meters is thus interpreted as the C1 TST interval.

One of the most prominent hardgrounds in the lower Wise Lake (Sinsinawa Member) is capped by a brown, organic-rich shale in some areas and is interpreted as the C1 maximum flooding surface to early HST deposit. Data from Fanton and Holmden (2007) indicate that Edenian-aged shales of the Wise Lake (although relatively thin) to show a pronounced upward increase in neodymium isotopic values showing once again a significant, even more pronounced, longer-term influx of Taconic-derived siliciclastic sediments (their I6 excursion) that maxes out below the Dubuque Formation in the Maysvillian. Thus, these shales are the extreme lateral equivalents of the Kope Formation and the Indian Castle Shales farther east. In the upper Mississippi Valley region, the C1 sequence, and similar scale sequences above, continues to show carbonate-dominated sequence architectures through the Cincinnatian.

Cincinnati Arch

In the Cincinnati Arch, the C1 sequence is represented by the upper Point Pleasant Formation and the overlying Kope Formation or the uppermost tongue of Tanglewood Member of the Lexington Limestone in the Clays Ferry Formation of Kentucky. Details of this depositional sequence have been well-studied and are discussed by McLaughlin and colleagues (2004) and McLaughlin and Brett (2004, 2007). It has been documented to contain abundant and intensely deformed seismite horizons and represents another pronounced episode of tectonic activity in the Taconic Foreland Basin and adjoining regions of the GACB.

As with the Steuben Formation in New York State, the upper Point Pleasant Formation is a widespread interval of crinoidal grainstone that signals relatively shallow, but transgressive conditions representing the last stand of the GACB. As in New York, the cross-bedded, coarse-grained shoal facies grade laterally into more distal basinal facies characterized by fine-grained turbiditic calcarenites–calcisiltites and interbedded black shales. These are very similar to facies found in the upper Dolgeville Formation of New York and the Milesburg Member of central Pennsylvania which are lateral equivalents.

The upper contact of the Pt. Pleasant shows an abrupt transition to a back-stepping succession of *Triarthrus*-bearing shales and argillaceous packstones of the Fulton Member of the Kope Formation and is interpreted as the late TST to condensed interval of the sequence. The sharp change then into a major shale-rich succession, of the basal Kope Formation is representative of the early HST facies. The early HST is indicated by development of dysoxic to anoxic conditions recording sediment starvation, followed by a substantial influx of siliciclastics into the foreland basin and across the Lexington Platform region.

Central Pennsylvania

Although the eastern Valley and Ridge succession had long been dominated by siliciclastic deposition east of the Adirondack Arch beginning in the M5B HST, the exposures in northwestern central Pennsylvania continued to show mixed siliciclastic and carbonate deposition through the end of the M6C sequence. The region was apparently a deepwater ramp coming off the shallower water environments of the Trenton Shelf to the north and may have been isolated at least partially by the submerged Adirondack Arch to the east. As in New York in the western Mohawk Valley, the final episode of carbonate-influenced deposition occurred in the C1 sequence as recognized here. The entire Coburn Formation represents a deep water turbidite/rhythmite succession, although prominent fine to medium-grained, crinoid-bearing calcarenites and occasional *Sowerbyella*-bearing skeletal packstones are recognized nearly to the contact with the Antes Shale. The increase in crinoids (albeit fragmented) in the uppermost Coleville Member suggest affinities for deepwater equivalents of the Steuben TST. Moreover the prominent change to black, dysoxic to anoxic, *Triarthrus*-bearing shales in the base of the Antes Formation, suggests this interval is the latest TST to earliest HST facies of the C1 sequence and is the lateral equivalent of the Kope Formation and Indian Castle Shales elsewhere.

DISCUSSION & CONCLUSIONS:

During the Late Ordovician (mid-late Mohawkian), the "Great American Carbonate Bank" of eastern Laurentia became increasingly segmented due to the impact of the Taconic Orogeny. Initially the GACB formed a broad, fairly uniform, shallow-water platform across much of eastern Laurentia; however, the platform was subsequently divided into a number of shallow water platforms/shelves separated by intracratonic troughs and foreland basin segments. As indicated in this study and those cited herein, New York's Trenton shelf to Taconic Foreland Basin and the Lexington Platform to Sebree Trough (contemporary platform to basin ramps) were formed synchronously during the M5A-M5B sequences. These study areas were separated by over 900 km and isolated from one another by the Pennsylvania Embayment, which itself was segregated into a number of topographically distinct features although distinctly different from New York or Kentucky.

It appears that both of these platform areas were isolated from cratonic interior areas by additional platforms including the Galena Platform, etc. The two platform-basinal ramp transects differed significantly in regards to their geographic orientation, paleoceanographic wind and current patterns, and their proximity to major active orogenic centers of the Late Ordovician. In addition, the tectonic settings of the two regions were also distinct. The Trenton Shelf, itself partially segmented by local block uplifts, was located immediately northwest of the actively subsiding Taconic Foreland Basin and portions of it have been argued to represent a forebulge. The Lexington Platform, itself comprised by fault-bounded localized uplifts, in turn was formed on the southeastern edge of the Sebree Trough. The Sebree Trough was also a fault bounded depression formed some 250-300 km inboard of the nearest plate margin. Thus the Lexington Platform – Sebree Trough pair has been considered as a peripheral forebulge to back-bulge basin respectively. The architecture of these features and their association with deep-rooted cratonic interior structures suggests they are ancestral features that became reactivated during Taconic orogenesis. Whether they are a forebulge-to-back-bulge basin respectively is still not clear. Nonetheless, as documented by Brett and colleagues (2004), given the distance and differences between the two paleogeographic areas, it is surprising that these areas show remarkably similar patterns of facies change both vertically and laterally during the Mohawkian sequences documented herein.

As suggested herein, the distribution of distinct facies at certain levels over wide areas is likely tied to variable, but widespread changes in climate and sea-level. As outlined in chapters 3, 4, & 6 and detailed in discussions of chapter 6, facies similarities and the magnitude of facies change during each coeval sequence was remarkably consistent over large areas of the GACB, especially throughout the M2-M4A sequences (late Ashbyan - late Turinian). During this time, shallow-water carbonate deposition overcame siliciclastic deposition across much of the GACB, even in the upper Mississippi Valley near the Transcontinental Arch during the early Tippecanoe Megasequence. Thereafter during the M4B to C1 sequences, although the similarity of facies between all regions became somewhat less pronounced due to tectonic evolution of the platform margin and craton interior, facies developed in both New York and Kentucky remained nearly indistinguishable during each respective sequence. This suggests that despite evidence for significant structural change to the GACB due to the Taconic Orogeny, the development and distribution of facies was at least partly controlled by sea-level and climate during this time.

For instance, as emphasized by Holland and Patzkowsky (1996), the occurrence of fenestral micrites, and other micrite-dominated facies in these sequences indicates widespread, relatively low-energy, shallow water environments (i.e. M2-M4A sequences). Given paleo-latitudes for the entire GACB region, and its relative isolation from deep water, open-ocean environments, the GACB was likely bathed by relatively warm water. These waters were likely supersaturated with respect to calcium carbonate and were favorable for both biotic production of micrites (algal and/or microbial) as well as abiotic precipitation of calcium carbonate as conditions were favorable (i.e. increased pH, lowered organic matter, etc.; Chave & Seuss, 1970).

Although micrite production continued through the Chatfieldian, the rate of production was significantly reduced during “Trenton Group” sequences. In all post-“Black River Group” sequences carbonate production was influenced by water depth and the influx of siliciclastic sediments that began to inundate platform margin areas, and indeed the entire platform.

Beginning immediately following deposition of the Deicke K-bentonite in the M4A sequence (in the Upper Mississippi Valley) and more regionally during the M5A sequence after deposition of the Millbrig and Elkport K-bentonites, orogen-derived siliciclastic sediments become more prevalent across the GACB.

Initially, an increase in shale-dominated deposition, and a reduction in carbonate, is observed in the Decorah Formation. This indicates a significant evolution in the physical geography and environments of the Hollandale Embayment-Wisconsin Arch area in the northernmost Upper Mississippi Valley after deposition of the Platteville carbonates (Emerson et al., 2004). Thereafter, beginning with the M5A sequence HST, thinly-bedded, rhythmic calcilutites and shales become important in all three primary study areas of the central and eastern GACB. With the significant contribution of micrite in these facies, carbonate production evidently continued in some areas, although storm activity and more intense paleoceanographic circulation within the GACB interior was involved in the redistribution and deposition of these micrite-rich facies in deeper water environments. For instance it is argued that the shallowest portions of the Trenton Shelf, north and west of the Taconic foredeep, supplied significant amounts of micrite to deeper water environments found 200 km or more to the south in central Pennsylvania. Thus, these carbonate-shale facies are poised between the production/deposition of carbonates and delivery of increased siliciclastics which has been attributed to periodic climatic fluctuations and/or changes in paleoceanographic conditions that become more distinct during particular sea-level phases. Nonetheless, it is apparent that these environmental conditions were pervasive over much of the eastern quarter of Laurentia and perhaps further prior to major shale and flysch progradation in the Edenian.

In addition to similar development in facies between study regions, this study has build on the work of Brett and colleagues (2004) to show that in addition to the six, third-order, Chatfieldian or "Trenton Sequences" (M5A-M6C), it is possible to recognize at least six, third-order sequences in the pre-Chatfieldian Ashbyan to Turinian interval (M1A –M4B). Using the constraints of K-bentonites, biostratigraphy, and other event correlations, this study shows that the number of sequences and component subsequences match relatively well between study areas, and especially between New York and Kentucky. In most instances, distinctive, detailed sedimentologic patterns and biostratigraphic composition of individual sequences appear to match very closely at meter- to decameter-scales. Sequences are most easily identified when they are considered across the entire gradient from shallow to deep water ramp environments. In the shallowest, most proximal portions of ramps, facies become much more amalgamated in the vicinity of erosion surfaces. Therefore boundaries of depositional sequences and systems tracts become challenging to recognize. Down-ramp, in more distal basinal settings, sequences tend to be dominated by greater percentages of dark shales with thinner, condensed carbonate interbeds. In these areas (including within the Pennsylvania Embayment, the Sebree Trough, and the Taconic Foredeep Basin), sequences and component cycles are much more subtle and again are challenging to recognize. Therefore, sequence correlations are most permissive and most easily identified when they are established using the most heterolithic portions of each transect. These are often developed in intermediate ramp sections. In these mid-ramp environments, amalgamation is less pervasive, condensation occurs but in distinct horizons, and shale-carbonate deposition is more or less predictably developed during different sea-level phases.

Despite some differences in sequence thickness, careful attention to detail and recognition of key surfaces and facies trends permits recognition and correlation of cycles across major environmental transitions on the ramp, both in the Lexington Platform–Sebree Trough (McLaughlin et al., 2004) and in the Trenton Platform–Taconic Foreland (Brett and Baird, 2002, Baird and Brett, 2002). These data support a strong, allocyclic, eustatic control on the development of third and fourth-order sequences (as well as on higher order cycles) in the study region. Moreover, when facies transitions are compared between successive sequences along each ramp transect, it is also possible to identify atypical patterns in facies development and cycle thicknesses, which in most cases are thought to be linked to autocyclic, tectonic-induced phenomena.

In contrast to Joy and colleagues (2004), who suggest that the major cycles documented in the Taconic Foreland are mostly tectonic in origin and asynchronous across the eastern GACB, this study provides evidence for a much more synchronous modification of facies during the onset of the Taconic Orogeny and a continued eustatic sea-level overprint on facies development through the Chatfieldian. However, through detailed analysis of sequences and sub-sequence components, both within and between study areas, it is clear that climate and tectonic signatures changed substantially within the approximately 10 million year time interval of this study. It appears that major changes occurred in rather rapid pulses with major environmental (and biotic transitions) occurring concomitantly. The first major pulse occurred in the M1B – early M2 sequences near the Chazy-Black River Group boundary with the significant reduction of siliciclastic sedimentation across the shelf, establishment of the immense "Black River lagoon," and formation of the Sevier Basin. The second major pulse occurred during the M4B through the M5A sequence (Black River –Trenton Boundary); a third pulse occurred during the M5C-M6A sequences (lowest Shermanian), and a final major pulse during the M6C-C1 sequences. Through each of these pulses, depositional sequences show small to moderate increases in rates of subsidence (and uplift) in both local to more widespread areas of the GACB along with a significant number of K-bentonite horizons, and an increased number of seismically deformed strata. All of these data indicate the increasingly important role of tectonics, not only on the platform margin, but across much of the GACB during the Chatfieldian to early Edenian.

As noted, after initial flooding of the GACB (M1A-M1B sequences) Turinian to early Chatfieldian sequences (M2-M5A) are characterized by remarkably uniform facies over relatively large areas (~200 km or greater) in Ontario–New York, Kentucky–Ohio, and central Pennsylvania. In distinct contrast, beginning in the lowest Shermanian (M5C), subsequent sequences show more abrupt lateral facies change as indicated by Brett and colleagues (2004). As documented in detail by Brett and Baird in New York (2002), and McLaughlin and colleagues (2004) in Kentucky, these later sequences often show abrupt transitions laterally across minimal distances (i.e. ~10-20 km). In the Mohawk Valley, cycles can be traced from pack- and grainstone facies deposited on fault-uplifted highs northwest of the Taconic foredeep into much darker graptolitic shales in areas of the Taconic foredeep basin to the southeast (Brett and Baird, 2002). Likewise, in Kentucky, sequences have been traced from peritidal fenestral micrite facies in the southern Lexington Platform to dark, graptolite-bearing shales in the Sebree Trough (sequences M6A and M6B; McLaughlin et al., 2004). In contrast in central Pennsylvania, gradients are much less distinct and depositional sequences are much more uniform in their appearance over wide areas although a general pattern of deepening is observed from north to south away from the Trenton Platform. The first sequence of the Edenian (sequence C1) shows a return to much more widespread facies with the rapid progradation of mud-dominated sediments across the eastern GACB (New York, Pennsylvania, and Kentucky).

Similar to the outcomes of Brett and colleagues (2004), this study shows the GACB region became increasingly partitioned by small subsiding basins and respective local highs. These basins and local highs often show evidence for inversion and high frequencies of soft-sediment deformation, especially in sequences with rapid facies change from proximal to distal ramps, (McLaughlin and Brett, 2002; 2004). These data suggest that changes in depositional facies are a direct result of movement on reactivated ancestral normal faults – some of which formed during the opening of the Iapetus Ocean and therefore pre-date the Late Ordovician. Interestingly, when compared between different parts of the foreland basin, the sequences that show the most evidence for tectonic influence occur synchronously (i.e. M6A, M6B, M6C, etc.). This strongly argues against a simple flexural bulge migration as suggested by some authors (see chapter 8).

Instead it is suggested that intervals of time with rapid lateral facies change and more intense syn-sedimentary deformation across the GACB record episodes of intense, oblique, compressional thrust-loading of the craton margin, with associated load readjustment along ancestral faults. In some cases, this readjustment takes place far from the tectonic load (see Rast et al., 1999; Ettensohn et al., 2002). This readjustment is argued to have occurred in both the Taconic Foreland and the Sebree Trough at approximately the same time (Brett et al., 2004). Moreover, the earliest of these "readjustments" are tentatively correlated herein to the first major pulses of coarser-grained, flysch-style siliciclastics (siltstones and fine sandstones), within the Martinsburg Formation of eastern Pennsylvania and the Stony Point Formation of eastern New York and western Vermont. In subsequent sequences and certainly by the C1 sequences, these eastern foredeep slope facies were incorporated into the frontal thrusts as the tectonic load intermittently pulsed westward.

As suggested previously by Brett and colleagues (2004), changes in depositional sequences associated with areas of uplift and subsidence, especially during the mid- to late Chatfieldian, indicate that tectonic processes became an increasingly important control in the development of Laurentian platform and basin architecture. Moreover, although these changes could have been produced by the complex interaction of multiple migrating flexural waves initiated by different collision centers at different times, it is important to note that tectonics did not completely overprint the pattern of climate and sea-level change that was so pervasive.

As shown herein, depositional sequences in the Late Ordovician (Chazy, Black River, and Trenton groups) have been identified and correlated between New York/Ontario, central Pennsylvania, and Kentucky/Ohio. Depositional sequence nomenclature herein expands the system first established by Holland and Patzkowsky (1996, 1998) for the Nashville Dome – Jessamine Dome region of Kentucky and adapted by Brett and colleagues (2004) for the New York/Ontario region. This paper also, for the first time, tentatively establishes a detailed correlation with depositional sequences described from the Late Ordovician of the Upper Mississippi Valley by Witzke and Bunker (1996). This contribution is important in that depositional sequences have now been constructed for the classic type localities of the early Late Ordovician and helps link important type localities from around the GACB platform. Moreover, this framework allows for an analysis of the timing and impact of the main phases of the Taconic orogeny and their relative impact on the GACB of eastern Laurentia.

Despite the historical conflicts of challenging methodological perspectives and emphases on local facies development, detailed comparisons documented in this study clearly indicate the importance of allocyclic controls on the development of sedimentary cycles. Moreover, important changes in sequence architecture can be attributed to pronounced local to regional modification of autocyclic phenomena, most notably topographic/bathymetric change. The salient conclusions of this paper can be summarized as follows.

(a) The Upper Ordovician (Ashbyan-Mohawkian–lower Cincinnatian) Chazy, Black River, Trenton and Utica Groups in the type area of New York State/Ontario can be divided into at least 13 third-order depositional sequences each having an average duration of approximately one million years. The distribution of these sequences and internal systems tracts indicate that the New York Platform/Adirondack Arch was a pronounced positive feature during deposition of sequences M1A, M1B and M2. Thereafter, sequences M3 and M4A record inundation of significant portions of the New York Platform (as well as other craton interior areas) and demonstrate maximum uniformity of facies across the region. Beginning with deposition of the M4B sequence approximately concurrent with deposition of the Millbrig K-bentonite, a number of significant changes in facies belts are noted and include pronounced uplift/subsidence of local blocks of the western and central Mohawk Valley (Adirondack Arch). Correlation of the M5A and M5B sequences from the Trenton Platform eastward into the Taconic Foredeep show a pronounced condensation of facies reflecting foundering of the carbonate platform in areas east of the Canajoharie Arch. Delivery of muds initiated during the M5A sequence; however, shale-dominated facies first become prominent in the central Mohawk Valley in the M6A sequence (lower Flat Creek Shales). Shale-dominated facies completely cover the Trenton Platform of New York by the C1 HST (Indian Castle Shales).

(b) In central Pennsylvania, strata deposited in the Pennsylvania Embayment region are also divided for the first time into at least 13 depositional sequences similar in scale to those recognized in New York. Facies developed in the embayment show evidence for high rates of subsidence near the platform margin, especially during the M1A-M1B sequences but are typically characterized by poorly fossiliferous restricted dolomitic limestone and dolostone facies. Similar to New York, beginning with deposition of the M2 sequence, and coincident with the prominent (but tentatively recognized) strontium isotopic excursion at the Ashbyan-Turinian boundary, the Pennsylvania Embayment shows substantially reduced subsidence rates. During the M3 to M4A sequence, the area shows evidence for uplift of the Adirondack Arch which eventually dissects the embayment into a western Valley and Ridge basin and an eastern Valley and Ridge basin (Cumberland Valley and areas east). Concurrent with the late M5A TST, shallow water carbonate production in the Pennsylvania Embayment becomes non-existent. It is replaced by transported (turbiditic-style) micrite and shale rhythmite facies of the Salona Formation (in the "western Valley and Ridge basin"). The "eastern Valley and Ridge basin" of south-central Pennsylvania, displays coarse siltstone to fine sandstone and shale facies that interfinger with carbonates in the lower Martinsburg Formation. Thereafter, sequences in the western Pennsylvania Embayment are characteristically distal (deep water) although evidently protected from coarse-siliciclastics that were trapped east of the submerged Adirondack Arch. Concurrent with the major progradation of carbonate-rich facies in the Dolgeville-Steuben interval of New York, (sequences M6C-C1), depositional sequences in the western basin show an increase in accumulation rates immediately prior to the onset of shale-dominated deposition of the Antes Shale. Carbonates in this interval were likely derived from the Trenton Shelf while increased thicknesses of shales evidently originated from the south and east across the still submerged Adirondack Arch. By the C1 sequence, the western basin, and the Adirondack Arch to its east, had entered the foredeep basin as it pulsed successively westward.

(c) As outlined by Brett and colleagues (2004) the equivalents of the upper Black River, Trenton, and Utica groups in Kentucky-Ohio can also be subdivided into sequences. In addition to the eight sequences recognized in the previous study, an additional five sequences are recognized herein. The lowest depositional sequences (M1A-M2) are recognized in the subsurface mostly through core analyses. It is evident that these depositional sequences were deposited on a highly irregular surface produced, in part, during the Knox Unconformity. It is also evident from preliminary analysis that the M1A-M1B sequences are relatively thin when compared to their counterparts elsewhere suggesting the region was relatively starved for both siliciclastic and carbonate deposition during this time. During the M2-M4A sequence, topographic expression becomes more subdued, and uniform, and carbonate accumulation rates increase and surpass rates of other regions, especially during the M4A sequence. It is important to note however, that the southeastern Jessamine Dome region does show evidence for increased restriction and development of dolomitic limestones and dolostone-rich facies during the M4A sequence suggesting at least some localized movement on Kentucky River faults. Subsequent to the M4A sequence, depositional sequences and their component systems tracts correlate well with those of New York on the basis of biostratigraphic and event stratigraphic criteria.

(d) As with the Nashville Dome, the New York Platform, the Pennsylvania Embayment, and the Jessamine Dome of Kentucky, Late Ordovician strata of the Upper Mississippi Valley region have also been subdivided into third-order depositional sequences based on a review (with minor modification) of transgressive-regressive cycles established by Witzke and Bunker (1996). Thus strata of the upper Mississippi Valley, beginning with the St. Peter Sandstone, and inclusive of the overlying Platteville, Decorah, and Dunleith Formations have also been integrated into the 13 sequence model proposed herein. Recognition of depositional sequences in the region is only preliminary and tentative, but a number of key correlations enable the connections to be made. Key facies changes are first observed, as elsewhere in the M2 sequence when carbonate deposition first dominates and subdues siliciclastic-dominated deposition of the M1A and M1B sequences. Carbonate-dominated deposition is significantly reduced late in the M4A to starved M4B sequences as the Decorah Shale facies was deposited. During deposition of the M5A HST, following the second major strontium isotopic excursion, siliciclastic sediments show Taconic-derived neodymium chemistry at about the same time as the Guttenberg Isotopic Carbon Excursion. Ensuing depositional sequences are typically carbonate-dominated and show evidence of shallow-water deposition as were evidently developed on the Galena Platform, although facies are suggested to show periodic cyclicity as reflected in changes in relative amounts of carbonate and shales as well as through neodymium chemistry (Fanton & Holmden, 2007). Interestingly, many major faunal incursion epiboles do occur and show remarkable similarity and timing to those of the Jessamine Dome and Trenton Platform although much more detailed analyses are needed to differentiate and distinguish these.

(e) In terms of sequence and component systems tracts, as suggested previously, depositional sequences are most similar beginning with the M2 sequence in the earliest Turinian through sequence M5A (M3-M4A are the most similar overall). Depositional facies are most uniform and most widespread during the Turinian to earliest Chatfieldian with a significant number of "time-restricted facies" and unique marker intervals now recognized in this interval (see chapter 6). Moreover, faunal patterns are highly comparable between most regions and indicate intermittent periods of restriction and more open-marine circulation. Thus these sequences likely reflect allocyclic, eustatic fluctuations and associated climatic changes during a period of major expansion in the "Black River Lagoon." This was likely tied to a period of increased sea-floor spreading rates that may have resulted in eruption of the large-scale Deicke and Millbrig K-bentonites in the M4A-M4B sequences respectively. Although fairly uniform conditions persisted for some time thereafter (through the M5B sequence in NY and Kentucky) substantive changes were initiated during the M5A sequence through the introduction of siliciclastics across the entire GACB, followed by significant topographic contrasts that initiated in eastern Laurentia during and after deposition of the M5B sequence.

(f) As suggested above, depositional sequences are still identifiable in late Chatfieldian (Shermanian) strata and are most comparable between New York and Kentucky as noted previously. Nonetheless these sequences, and especially the M5C, M6A, M6B, and M6C, show substantially more variability in local areas than do older sequences. Most often these are manifest by synchronous, abrupt lateral facies change, and evidence for synsedimentary deformation across widespread areas (including in all three study areas). It remains to be seen if synsedimentary deformation/faulting events can be recognized in the Upper Mississippi Valley region coincident with those recognized elsewhere. Nonetheless, given their recognition in the Nashville Dome, the Jessamine Dome, the Pennsylvania Embayment, and the New York Platform, it is highly likely that these imply tectonism as a major control, not only on the architecture of localized regions of the Late Ordovician carbonate platforms of eastern Laurentia, but also on their final demise during the earliest Edenian (C1 sequence).

CHAPTER 8: Timing and Evolution of the Taconic Orogeny: Plate Tectonic Reorganization of eastern Laurentia during the Late Ordovician

ABSTRACT

As shown through detailed comparisons of thirteen Late Ordovician third-order depositional sequences, the demise of the "Great American Carbonate Bank" (GACB) of eastern Laurentia occurred simultaneous to the onset of the Taconic Orogeny and its multiple tectophases. Two distinct foreland basin-forming events are developed during the 10 million year interval (Ashbyan-Mohawkian stages), and document an active margin undergoing orogenesis. The first event, the Blountian tectophase, was small in scale, both temporally and spatially, while the later tectophase, the Vermontian, was much more substantial, impacted a much greater area, and persisted for a much greater length of time. Both foreland basins are documented to contain carbonate to shale to flysch to molasse wedges similar to other orogenic events elsewhere. Nonetheless, they are peculiar in multiple respects. Although the pattern of sedimentary fill is well known, this study seeks to resolve the spatio-temporal relationships of these foreland basin segments and seeks to relate these to patterns of uplift and subsidence observed throughout the leading edge of the Taconic foreland complex.

Using a refined sequence-stratigraphic model (Chapter 7) integrated with previous sedimentary provenance studies, the timing of basin fill and sedimentologic change in the cratonward portions of the foreland are herein documented. Widespread sea-level change exerted a strong influence on the deposition of sediments. Nonetheless, sequence architecture reveals patterns of localized to more regional facies change reflecting tectonic influences that correspond to events in the foredeep and hinterland. Given modeling constraints from: 1) the modern Timor-Banda Orogen; 2) numerical modeling of foreland basin load structures; and 3) chemostratigraphic data (documented in Chapter 6); it is suggested that differences in scale and

duration of the Blountian and Vermontian tectophases are not analogous and have important implications for the evolution of eastern North America at this time.

Through the synthesis of a multitude of data accumulated in the quarter decade since the last tectonic models were produced, an evolved view of Late Ordovician orogenesis and sea-level change is proposed here. Based on strontium isotopic evidence, a plausible correlation is tentatively outlined linking multiple pulses of sea-level change across the GACB with tectonic loading/basin subsidence episodes and with changes in the rate of seafloor spreading. These data provide evidence in support of a change in plate tectonic architecture of eastern Laurentia in the latest Turinian to earliest Chatfieldian stages. As such, the two primary phases of the Taconic Orogeny are herein considered to represent distinct phases of a single protracted collisional event (as opposed to analogous collisional events).

These tectophases are separated in time by a major plate tectonic reorganization that is recorded on the margin of North America by buoyant rebound and restriction of the GACB platform during the middle Turinian. This reorganization may have involved slab break-off, subduction reversal, and changes in the rate of sea-floor spreading as suggested by new strontium isotopic excursions. Slab break-off is followed in the late Turinian to earliest Chatfieldian by voluminous volcanic eruptions resulting from the rapid melting of the detached east-dipping subducted slab. By the middle Chatfieldian (early Shermanian stage), west-dipping subduction had initiated.

Recorded by a prominent change in sedimentary provenance, it is suggested that westward-directed telescoping of thrust sheets in the hinterland (accretionary wedge- forearc basin-volcanic arc) renewed development of the frontal wedge load. This resulted in a much more substantial foreland basin filled with a much more widespread dark shale succession (Utica

Group Shales) compared to the earlier tectophase. The onset of major shale deposition was coincident with development of the medial Trenton Group and its equivalents on the foreland basin margin. Maximum crustal loading of the Vermontian tectophase was completed in the earliest Cincinnatian coincident with deposition of the C1 depositional sequence (Indian Castle/Antes Shales of New York and Central Pennsylvania and the Kope Formation of the Ohio Valley). Hence as proposed here, closure of the forearc basin in the hinterland, development of a pronounced frontal wedge-load, is not only coincident with the last stand of carbonate deposition on the GACB at the end of the Mohawkian – but is responsible for its demise.

INTRODUCTION

The terminal phase of the GACB occurred inboard of a tectonically active margin undergoing orogenesis and subduction (Bird & Dewey; 1970; Rowley & Kidd, 1981; Shanmugam & Lash, 1982; Quinlan & Beaumont, 1984). The GACB itself was impacted by orogenesis well before its final demise at the end of the Mohawkian. Since the work of Rowley & Kidd (1981), Shanmugam and Lash (1982), and even earlier workers (Ulrich & Butts; as reported by Wilmarth, 1938; Kay, 1943; Rodgers 1971), it has been noted that sedimentary successions recording the onset of tectonic subsidence on the eastern margin of the GACB were of different biostratigraphic ages. Moreover, deepening and filling events occurred first in the south and then in the north. The southern phase, referred to as the Blountian tectophase, the Vermontian tectophase, was relatively small in scale both temporally and spatially. The second more northerly Vermontian phase was much more substantial and impacted a much greater area than the earlier tectophase and persisted for a much greater length of time (Chapter 6, **figure 2**).

Previous bio- and lithostratigraphic studies have addressed the architecture of sedimentary successions within and proximal to sub-basins produced during the Taconic

Orogeny (Rickard & Fisher, 1973; Fisher, 1977; Shanmugam & Walker, 1980; Shanmugam & Lash, 1982; Drake et. al, 1989; Lehmann et. al, 1994; Bergström & Mitchell, 1986; Ettensohn, 1991; 1994; Finney et al., 1996). Hence, the pattern of basin fill is relatively well-known and little question remains for the characterization of Blountian and Vermontian tectophase successions as representing standard flysch to molasse basin-fill wedges similar to those documented from the Devonian Catskill wedge or the classic Miocene Molasse wedge of the northern Alps (Sinclair, 1997; Garfunkel & Greiling, 2002). Thus, until now these tectophases have been considered analogous, yet distinct in their timing.

Notwithstanding, however, is the fact that very little was known about the relationships between these basins. With the refined chronology afforded by this investigation, the timing of basin filling and associated sedimentologic changes in the cratonward portions of the foreland basin are now becoming clearer. Sequence stratigraphic studies (as discussed in Chapter 7) have shown that prior to and contemporaneous with basin initiation, carbonate-dominated successions demonstrate several scales of sedimentary cyclicity. Thus, widespread sea-level changes exerted a strong control on the distribution of sediments. Moreover, within these sequence and cycle packages it is possible to discern patterns of localized to more regional facies change reflecting tectonic influences within the Taconic foreland basin and its related sub-basins (McLaughlin & Brett, 2004; Ettensohn et al., 2002b, 2004). Although fairly well-documented, especially in the Jessamine Dome region, it is still unclear as to how the timing of these eustasy controlled and tectonically-influenced packages correspond to events in the foredeep and hinterland.

The refined chrono- and event stratigraphic framework, established in this dissertation for the GACB (Chapter 2 and 7), has been integrated with data from sedimentary provenance studies. Herein these data are used to provide empirical constraints for the development of a new, more conciliatory model for the evolution of the Taconic Orogeny. Given constraints from: 1) the modern Timor-Banda Orogen; 2) numerical modeling of foreland basin load structures; and 3) chemostratigraphic data (documented in Chapter 6); it is suggested that differences in scale and duration of the Blountian and Vermontian tectophases record a pronounced reorganization in plate tectonic architecture of eastern Laurentia in the latest Turinian to earliest Chatfieldian, that likely involved slab break-off, subduction reversal and changes in the rate of sea-floor spreading. This plate tectonic reorganization, in turn, had major implications for the paleogeography of eastern Laurentia, and likely climate change leading into the end Ordovician.

PERIPHERAL-TYPE FORELAND BASIN ARCHITECTURE & DYNAMIC BEHAVIOR

Peripheral-type foreland basins (Dickinson, 1974) were distinguished as plate margin basins developed on a tectonic plate that undergoes subduction at the plate boundary. The formation of peripheral foreland basins occurs due to the deformation and flexure of the passive margin during collisional events and thrust loading of the plate margin. The resulting foreland basin system (figure 1) is composed of multiple sub-basins that may migrate through time as a

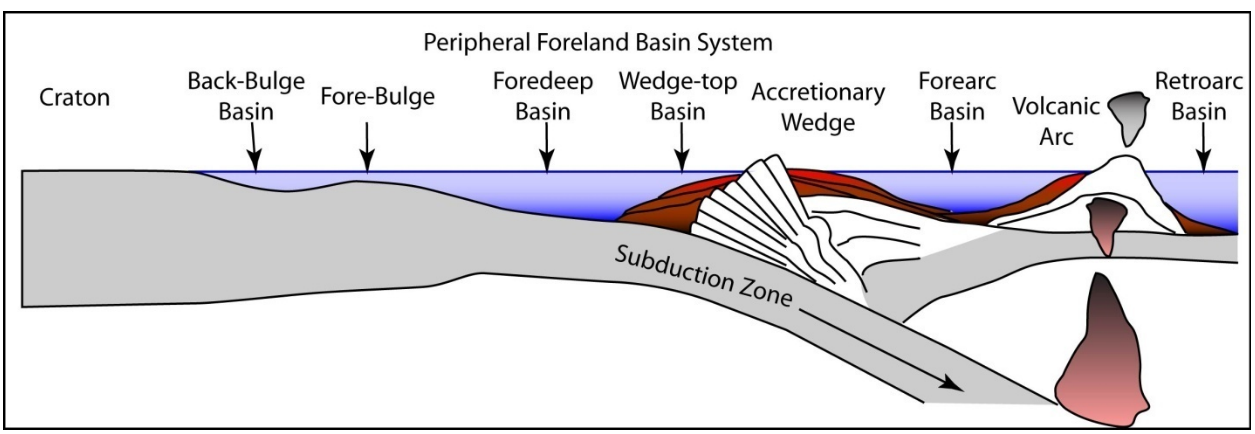


Figure 1: A typical foreland basin system, after DeCelles & Giles (1996), may be composed of a back bulge basin, forebulge, foredeep basin, and wedge-top basin that sits above the submerged portions of the accretionary prism. Further toward the hinterland, additional features include a forearc basin, volcanic arc, and a retroarc basin.

function of continued loading, flexure, and infilling of sediment derived primarily from the orogenic hinterland (accretionary wedge and volcanic arc; DeCelles & Giles, 1996).

Architecturally, it is clear that tectonic loading (i.e. growth of an accretionary wedge) along the plate margin is responsible for foreland basin subsidence in the immediate vicinity of the load (Price, 1973; Beaumont, 1981; Jordan, 1981; Lyon-Caen & Molnar, 1985; Stockmal et al., 1986; Flemings & Jordan, 1989; Sinclair et al., 1991; Watts, 1992). Theoretical modeling, assuming visco-elastic, and elastic rheological parameters, has shown that the formation of a flexural foredeep basin and a forebulge in front of a tectonic load depends on the properties of the subducting plate, and the size and shape of the load applied to the edge of the subducting plate (Garfunkel & Greiling, 2002). One outcome of these studies suggests that regardless of the shape of the load structure (accretionary prism/volcanic arc in the case of peripheral foreland basins) on the subducting plate, it is primarily the frontal 150 km of the load that influences the flexure of the foreland (Garfunkel & Greiling, 1996, 2002).

These studies have also shown that the width and depth of the foredeep basin depends on the taper angle of the load (roughly the height-to-width ratio of the accretionary prism), and on the flexural parameters of the crust being subducted. For instance, fairly young, warm crust, although somewhat buoyant, often deforms rapidly producing a fairly narrow foredeep basin (on the order of ~100 km). In contrast, older, cooler crust that is significantly denser tends to deform less producing a much broader foredeep (~200 km). Likewise, if the taper angle of the load is fairly low (i.e. the load is spread out over a broader region: fewer stacked thrust sheets with overall low relief) the foredeep tends to be shallow. Whereas, if the taper angle is high (i.e. more of the load is pushed to the front of the wedge: more stacked thrust sheets with greater elevation) the associated basin tends to be deeper.

Distal to the emplaced load, foreland basin evolution proceeds through a variety of topographic changes including uplift, and subsidence afforded by both compression and extension. As modeled theoretically (Jordan, 1981; Karner & Watts, 1983; Lyon-Caen & Molnar, 1985; Watts, 1992), and shown empirically in the Banda Orogen (**figure 2**), the main

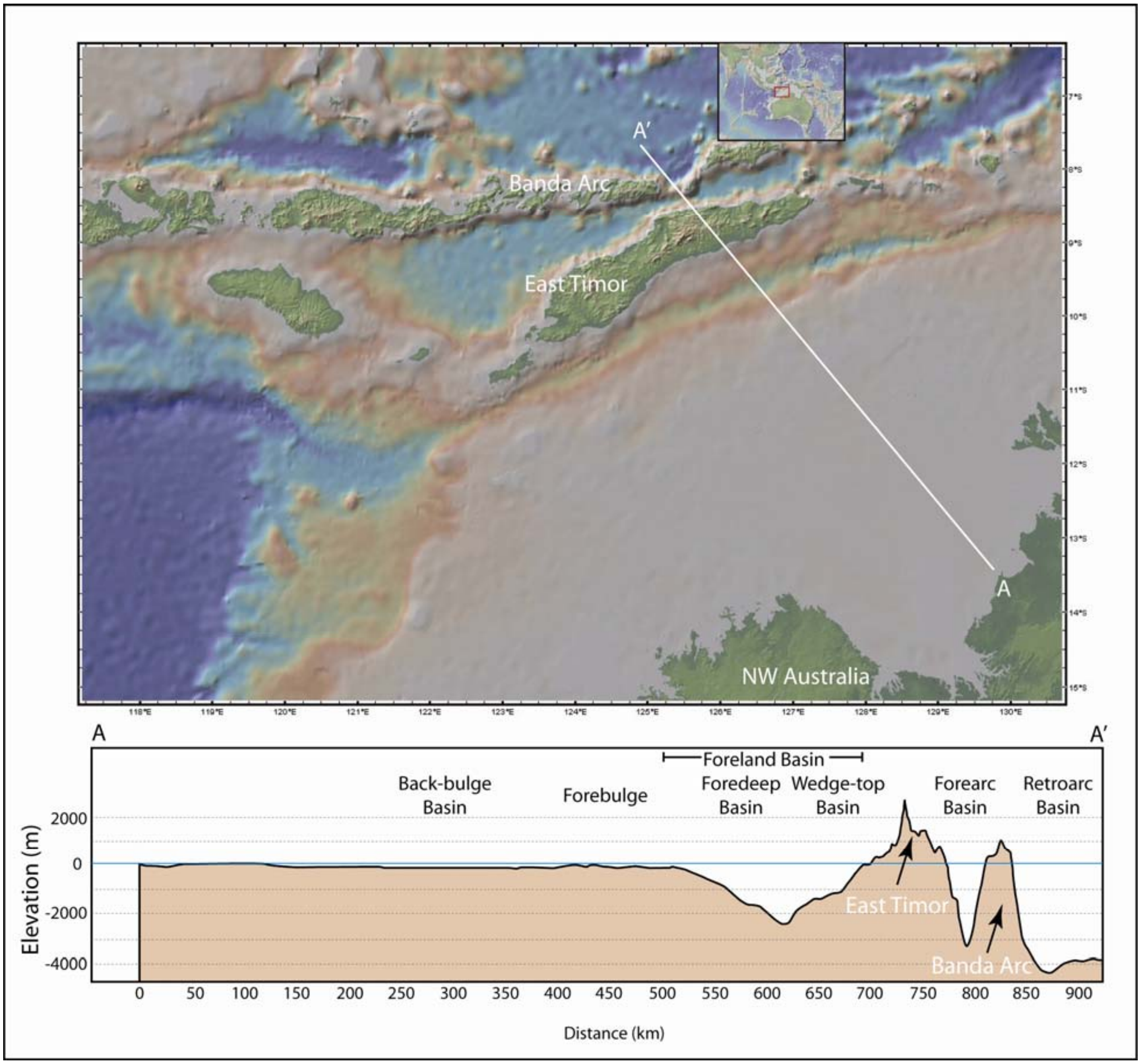


Figure 2: Topographic profile of the modern Banda-Timor-Sahul Shelf area of northwest Australia showing the architecture of the subduction zone and associated peripheral foreland basin. The foreland basin is developed on top of the north-dipping subduction zone of the Indo-Australian Plate below the overriding Eurasian Plate. Profile and map modified from graphics produced using GeoMapApp software.

uplift typically occurs some 100 to 200 km from the foredeep-frontal load margin. The uplift, referred to as a peripheral bulge, can be manifested as a fairly broad uplift equal to the width of the foredeep (Garfunkel & Greiling, 2002). Uplift can be accommodated by ancestral fault block reactivation, and formation of narrow horst block uplifts and intervening grabens (**figure 3**). As

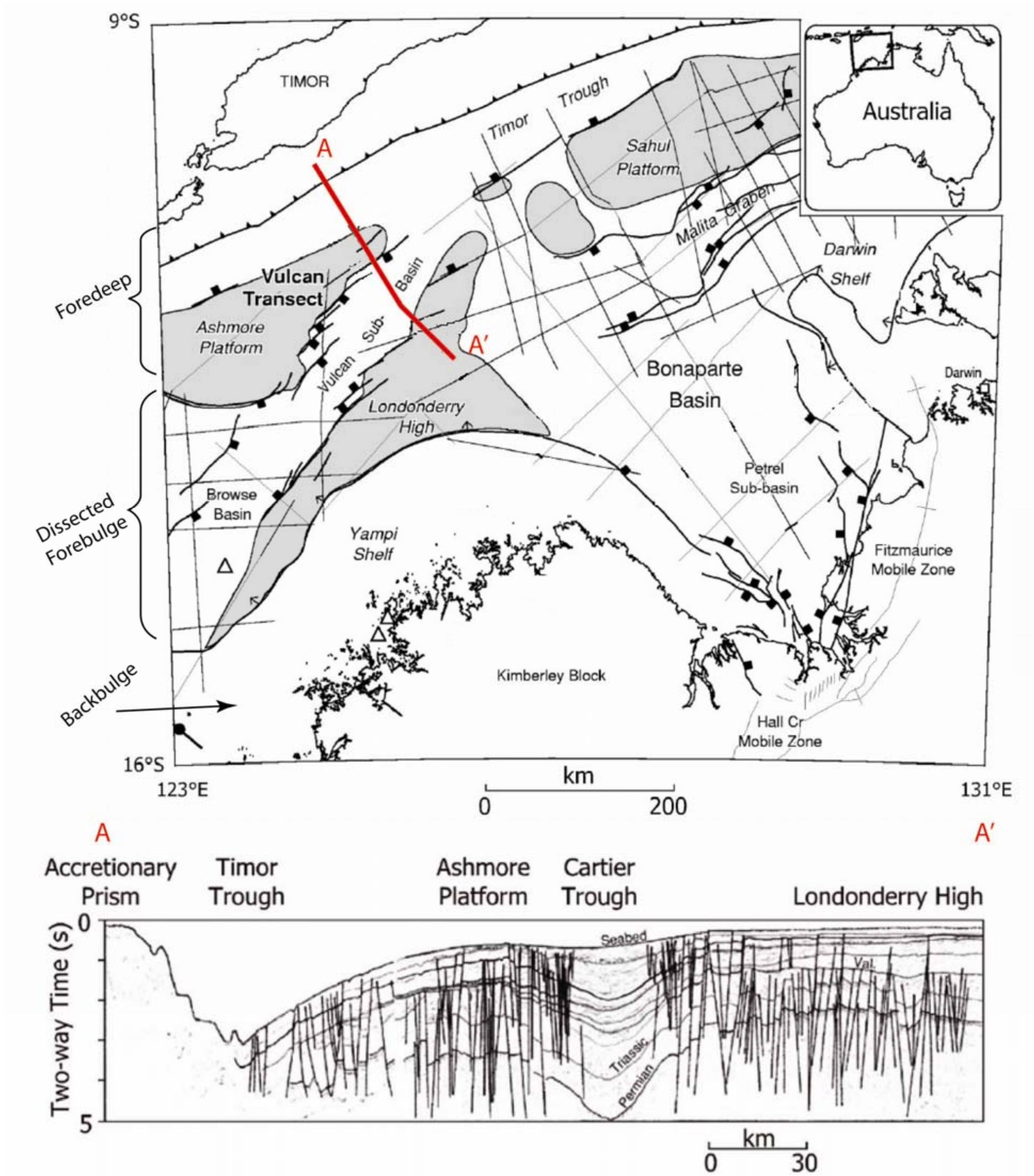


Figure 3: Map and composite cross-section from the Timor Trough southward onto the Australian continental shelf. As shown, the region between the foredeep (Timor Trough) and the forebulge is intensely faulted and dissected. Movement on these deep seated faults is argued to accommodate much of the flexure associated with the collision of the Timor Accretionary Prism (figures modified from Shuster et al., 1998; and Petkovic et al., 2000).

documented in modern foreland basins, it is known that these uplift and subsidence patterns can be locally influenced by ancestral faults that can be reactivated in advance of the foredeep (Petkovic et al., 1999). For the modern Timor-Banda Orogen, a modern analog for the Taconic Orogeny, Audley- Charles (2004) has shown that the width of the Banda forearc basin varies from ~50 to 100 km in width, while the accretionary wedge (East Timor) can range up to 150 km in width with approximately half forming the submerged wedge-top basin. The foredeep basin (Timor Trough-Ashmore Platform-Cartier Trough) through the forebulge (Londonderry High), in turn, is typically less than 200 km in width and the forebulge (i.e. Londonderry High) can range from 200 to 300 km in width. The back bulge region is variable in width, and depends on sea-level. As noted by Petkovic and colleagues (2000), tectonic activity (fault movement) has been noted up to 400 km or more from the foredeep along the cratonward edge of the forebulge. Peripheral bulges, once formed, have been postulated to either: 1) remain stationary until they are over-thrust (Shanmugam & Lash, 1982), or until the foredeep basin is filled (Holland & Patzkowsky, 1997); 2) migrate toward the craton as the over-thrust is pushed cratonward (Quinlan & Beaumont, 1984); or 3) migrate toward the load as the foreland is depressed (Jordan, 1981; Jordan & Flemings, 1991) (**figure 4**). Recent evidence for flexural migration of the forebulge in the modern Banda orogen is lacking as subduction of the Australian margin has failed and collisional stress is now accommodated along back thrusts developed in the overriding southeast Asian plate (i.e. along the Wetar and Flores thrusts; Audley-Charles, 2004) and in the accretionary wedge-arc region that appears to be uplifting but not migrating (Kaneko et al., 2007). Historical evidence from the same region suggests, however, that migration of the foredeep-peripheral bulge did occur as reflected by cratonward migration of the Australian shelf edge.

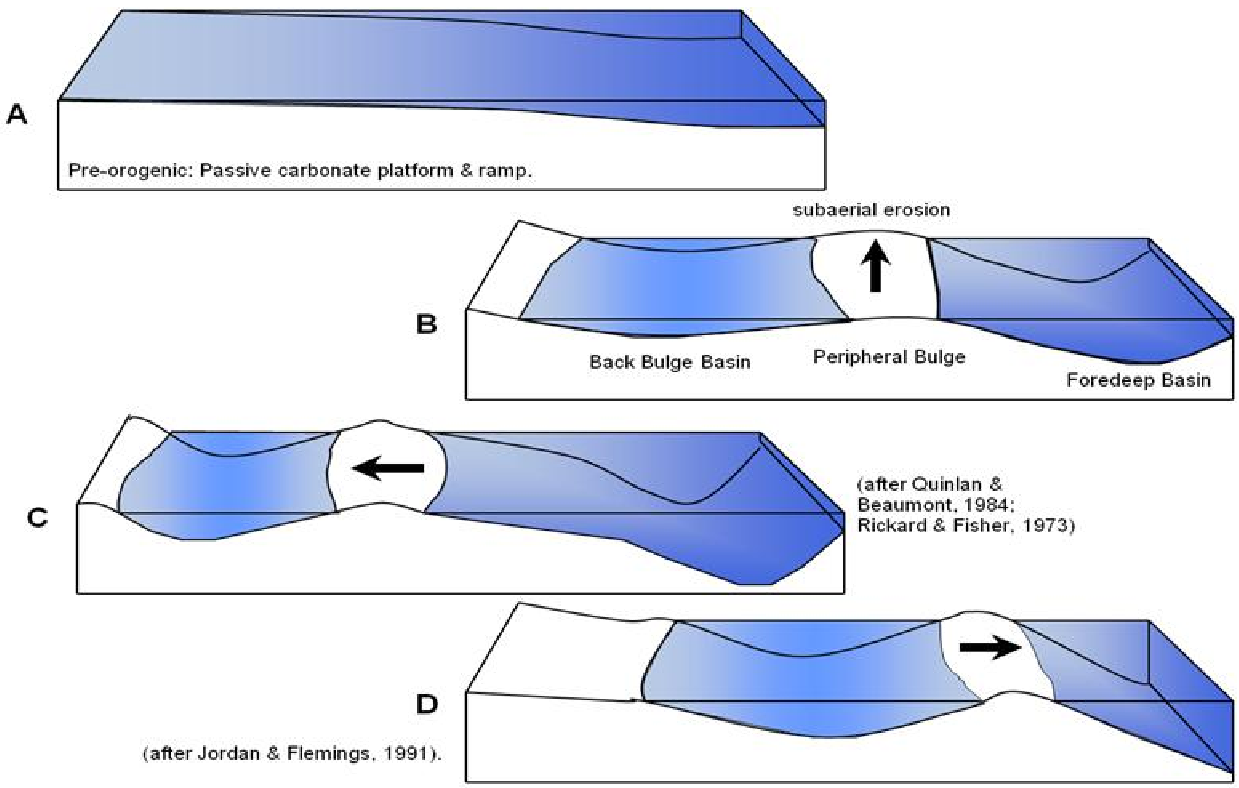


Figure 4: Theoretical Models for modification of continental margins during the onset of foreland basin development. A: pre-orogenic configuration of a passive margin, B: stationary peripheral bulge and associated sub-basins, C: cratonward migration of the forebulge, and D: migration of the forebulge toward the hinterland.

During the latest Miocene (5mya) the continental shelf edge was located some 150-200 km northwest of its present location. The Miocene shelf-edge is thus buried within the floor of the Timor Trough, and portions of the slope and rise were incorporated into the frontal edge of the accretionary prism before subduction was jammed (Audley-Charles, 2004; Norvick, 1979). Thus, at least for the modern Banda Orogen, it appears that forebulge migration subscribes to scenario two (C) above (after Quinlan & Beaumont, 1984). Nonetheless, it is still important to note that ancestral faults can be reactivated across the foreland basin system during loading and associated faulting has the capacity to produce numerous synsedimentary features across the foreland basin system. These include seismically deformed horizons, over-steepened gradients, etc., that signal major events in the hinterland even though subduction may have ceased (Bradley

& Kusky, 1986; Bradley & Kidd, 1991). Moreover, localized, fault-bounded, uplifts may undergo inversion during flexure, a scenario that is also suspected to have occurred in the Ordovician of eastern North American in both the Sevier Basin of Georgia and Tennessee and the Martinsburg Basin of central New York (Bayona & Thomas, 2003; Mitchell & Jacobi, 2002).

HISTORICAL MODELS & NEW EMPIRICAL CONSTRAINTS FOR THE EVOLUTION OF THE TACONIC OROGENY

Rowley and Kidd (1981) significantly updated previous models for the evolution of the Taconic Orogeny when they coupled available sedimentologic and paleontologic data from the foreland to interpret the structural evolution and timing of the Taconic Orogeny as it developed in New England. Their model shows the development of the Taconic foreland basin system beginning in the earliest Late Ordovician (Ashbyan Stage; **figure 5**). In this model, the Chazy-Black River unconformity (post-Chazy unconformity) is hypothesized to represent the first development of the foreland basin with uplift of the forebulge. In the foredeep basin, the event was thought to be recorded by the deposition of Indian River red muds on the distal slope and rise. These were presumed to have been derived from the uplifted bulge. Further toward the orogenic center (Ammonoosuc Arc), ocean floor sediments and ophiolitic materials were scraped off the Laurentian plate as it went into the east-dipping subduction zone. The accumulation of these materials into an accretionary wedge provided the load for the depression of the foredeep basin, as well as a new supply of sediment (flysch) that began to enter the foreland basin in the Turinian. By the end of the Chatfieldian, much of the Taconic allochthons were emplaced, including the Champlain thrust, the Taconic thrust, etc., and significant volumes of siliciclastic sediment derived from the volcanic arc prograded rapidly across the foreland basin and into interior portions of the Laurentian craton. In this model, significant progress was made in terms of explaining the pattern and timing of flysch development in the foredeep and wedge top basins, however, several important delinquencies arise.

First, the recognition of the Chazy-Black River unconformity, although plausibly linked to later, local accentuation of the forebulge, neglects recognition of the earlier and much larger-

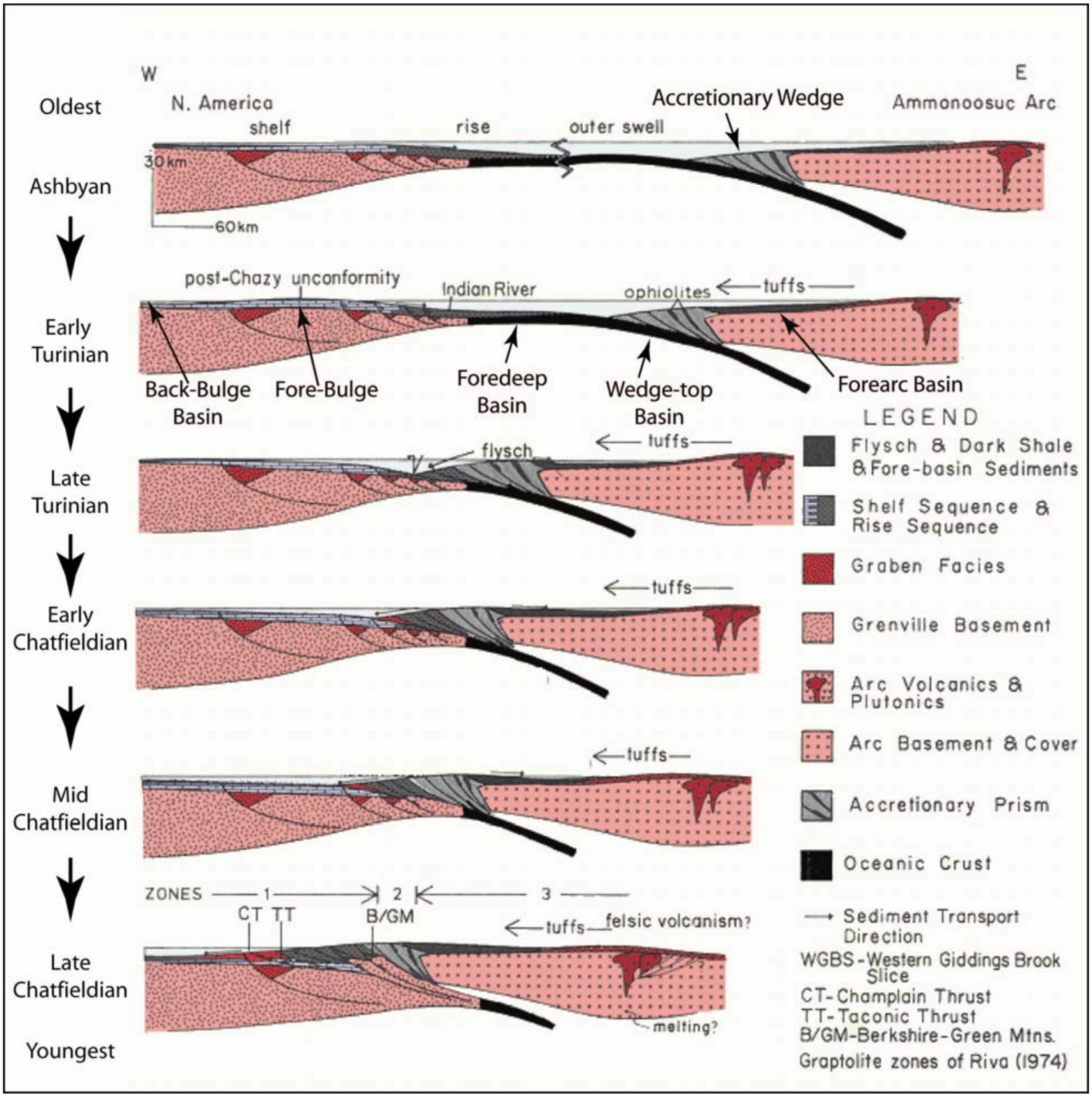


Figure 5: Schematic diagram showing the evolution of the Taconic Orogeny through time in western New England (New York, Vermont, and New Hampshire) coincident with deposition of the Chazy, Black River and Trenton Groups (Ashbyan to Chatfieldian). Figure modified and adapted from Rowley & Kidd, 1981.

scale Knox Unconformity and fault activity prior to and during deposition of Chazy Group rocks (see Chapter 3). Thus, if a forebulge developed in the region, associated with reactivation of ancestral basement faults, it appears to have occurred much earlier than the earliest Turinian as suggested in the Rowley and Kidd model. Furthermore, the Chazy-Black River unconformity appears to be relatively localized in its extent in New York/Ontario, although it is more prominent and persisted longer to the south into eastern Pennsylvania (as indicated by Lash & Drake, 1984; see below). Moreover, it is now shown that the Indian River red mudstones (supposedly derived from the uplifted bulge) are likely much older than assumed in this model. As such, the Indian River slates may actually record the position of the forebulge on the deep sea floor as it moved cratonward during the pre-Knox highstand (Landing et al., 2003; pers. com.).

Secondly, as pointed out by Karabinos and colleagues (1998), much of the volcanism associated with the Ammonoosuc/Bronson Hill Arc is younger than most deformation in the Taconic Orogen and reflects the onset of volcanism east of the Shelburne Falls Arc (**figure 6**). These data indicate the presence of a volcanic arc not accounted for in the Rowley and Kidd (1981) model. Apparently, the Shelburne Falls Arc was formed from partial melting of the subducted Laurentian plate well off the coast of eastern Laurentia in earlier Ordovician time and was coincident with volcanism and deformation identified in terranes to the south (see figure 11, Chapter 1). By the late Turinian, it appears that little new volcanic activity took place in the Shelburne Falls Arc. Nonetheless, given Turinian-aged deep-water flysch off the Laurentian coast, and in the Sevier Basin (Blountian tectophase), it is clear that accretionary wedge development had occurred and separated Laurentia from the Shelburne Falls Arc. Subsequently evidence for a surge in volcanism some 50-100 km or more to the east of the Shelburne Falls Arc (i.e. the Bronson Hill/Ammonoosuc Arc) indicates a pronounced change in the volcanic front during the early Chatfieldian. Karabinos and colleagues (1998) postulated the eastward shift in volcanism as the result of subduction reversal that took place about 454 mya (**figure 7**). Given

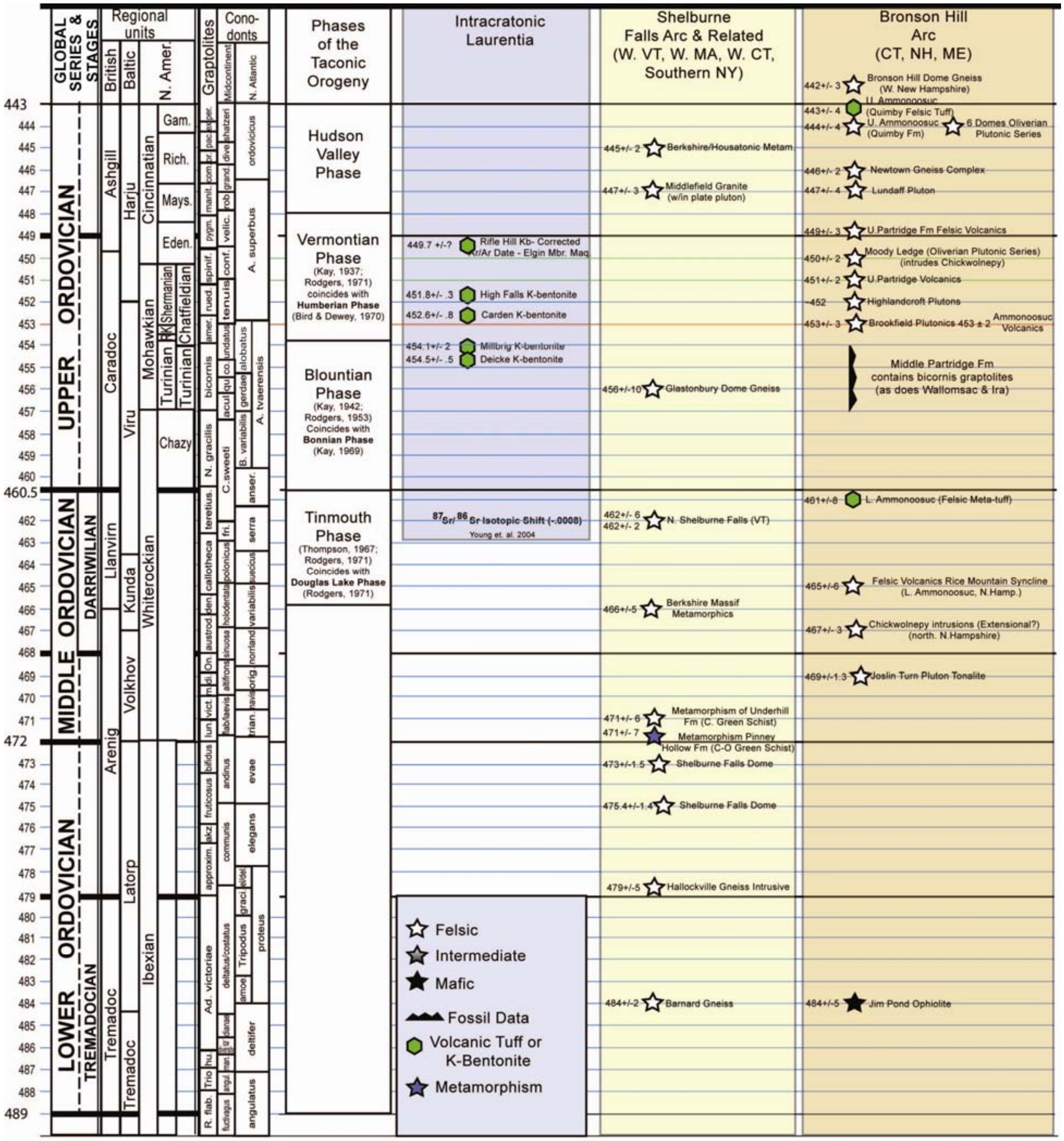


Figure 6: Chronology and absolute dates for volcanism and collision related events in the type Taconic area of New England. Radiometric dates taken primarily from Hollocher et al., (2002) and references therein. As shown, significant felsic volcanism in the Bronson Hill/Ammonoosuc Arc occurred immediately following the large-scale eruptions of the Deicke and Millbrig K-bentonites in the Late Ordovician well to the east of the Shelburne Falls arc that had formed mostly prior to the Blountian tectophase.

this age, this reversal coincides with the pronounced Turinian-Chatfieldian unconformity across the GACB within the proximal foreland basin. As suggested by the Rowley and Kidd (1981) model, the latest Turinian to Chatfieldian, in turn, saw the thrusting of the accretionary wedge up the continental slope and rise and onto the New York Promontory passive margin sequence (**see figure 5, figure 8**).

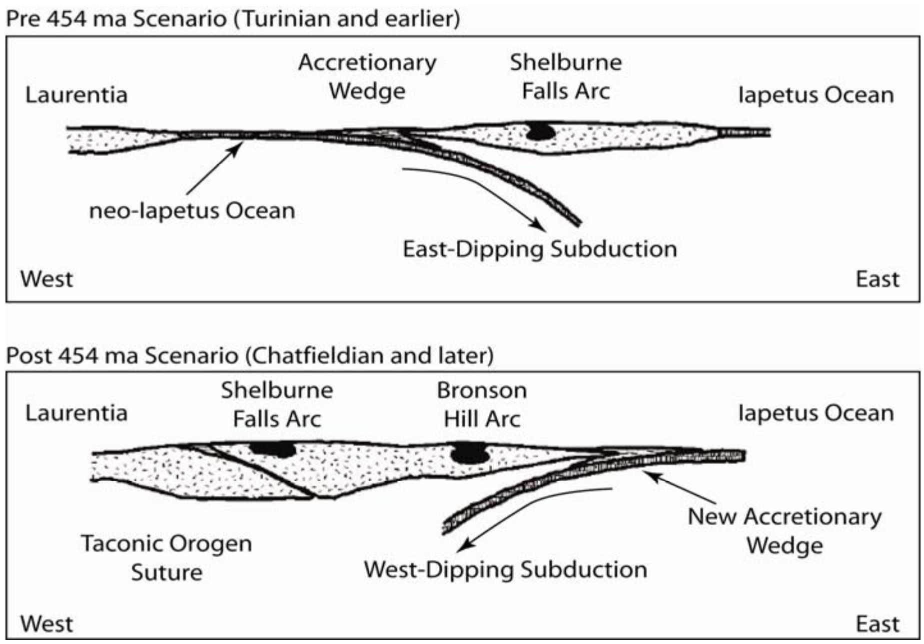


Figure 7: Schematic cross-section of the subduction reversal model proposed by Karabinos and colleagues (1998). Subduction reversal is thought to have occurred at about 454 ma at about the time the Millbrig K-bentonite was erupted.

As shown in more detail by Lash and Drake (1984), the occurrence of the "Black River hiatus" during the Turinian recorded accentuation of the peripheral bulge (scenario 1, or B from figure 4 above) that persisted for much of the Turinian. The uplift, rather than forming as a consequence of flexural wave migration, was interpreted by these authors to remain stationary due to aborted subduction of the buoyant continental shelf. Coincident with failed subduction of the GACB margin in the latest Turinian to early Chatfieldian, normal faulting appears to have reactivated basement faults resulting in subsidence of the continental shelf margin resulting in the formation of the foredeep. Moreover, thrust faulting (under-plating at the front of the wedge

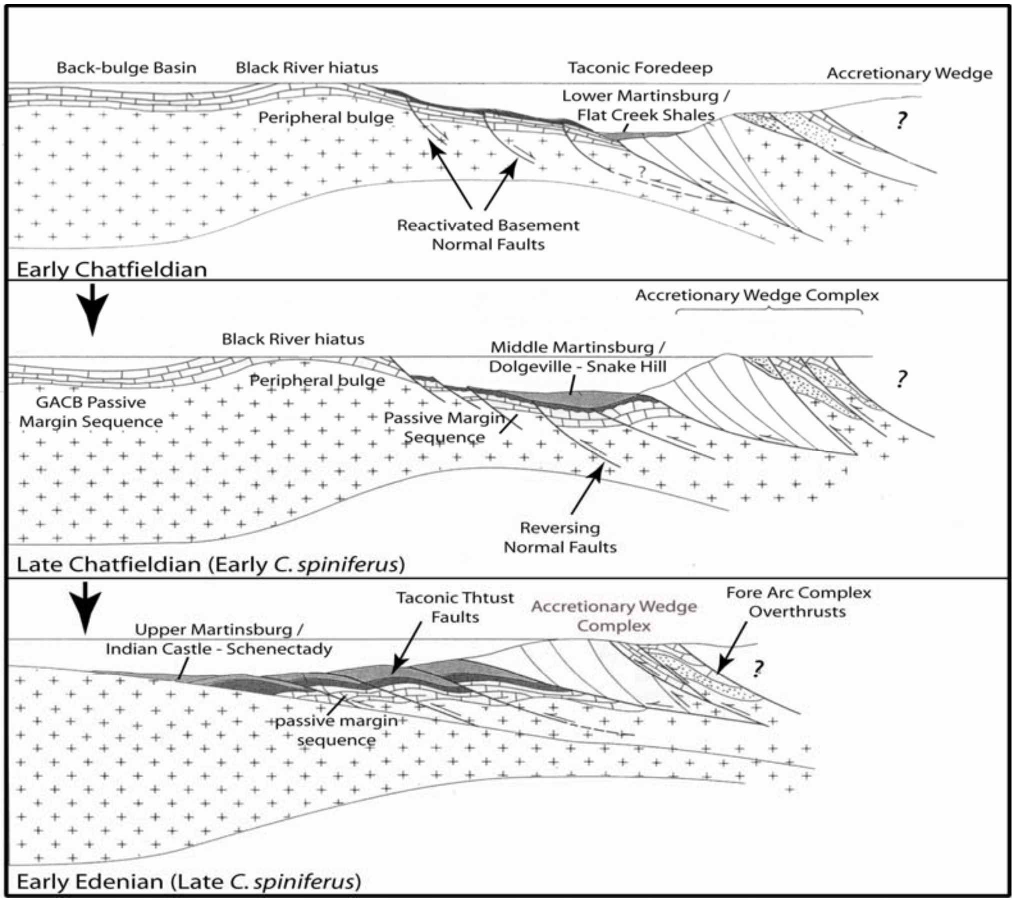


Figure 8: Evolution of the Martinsburg Foreland Basin as proposed by Lash and Drake (1984; updated to the stratigraphy and chronology used herein), and associated sedimentary and structural developments across the eastern portion of the Martinsburg Basin during the Vermontian Tectophase. In this model, the accretionary wedge complex is documented to show evidence of shortening and thickening from the east toward the forearc-volcanic arc.

and over-plating in the eastern-most wedge; Bosworth et al., 1984) through the Chatfieldian enabled uplift and compression of the accretionary prism, perhaps by as much as 50-60 km or more (Bosworth and colleagues, 1984). This accommodated the large-scale subsidence and influx of siliciclastics (Martinsburg Formation) into the foredeep. By the Edenian (late *C. spiniferus*-zone), Lash and Drake (1984) argued that much of the Hamburg Klippe, and other Taconic-type klippen, were in place, tectonic movement had slowed significantly (if not ceased), and basin-filling materials were becoming more mature (quartz-dominated).

Like Karabinos and colleagues (1998), Lash and Drake (1984) had already suggested a subduction reversal outboard of the accretionary wedge to the east of the volcanic arc/microcontinent terranes (i.e. Manhattan Prong, Shelburne Falls Arc, Baltimore Terrane, etc.; see Chapter 1: figure 11; **figure 9**). This reversal was argued by them to have occurred in the early

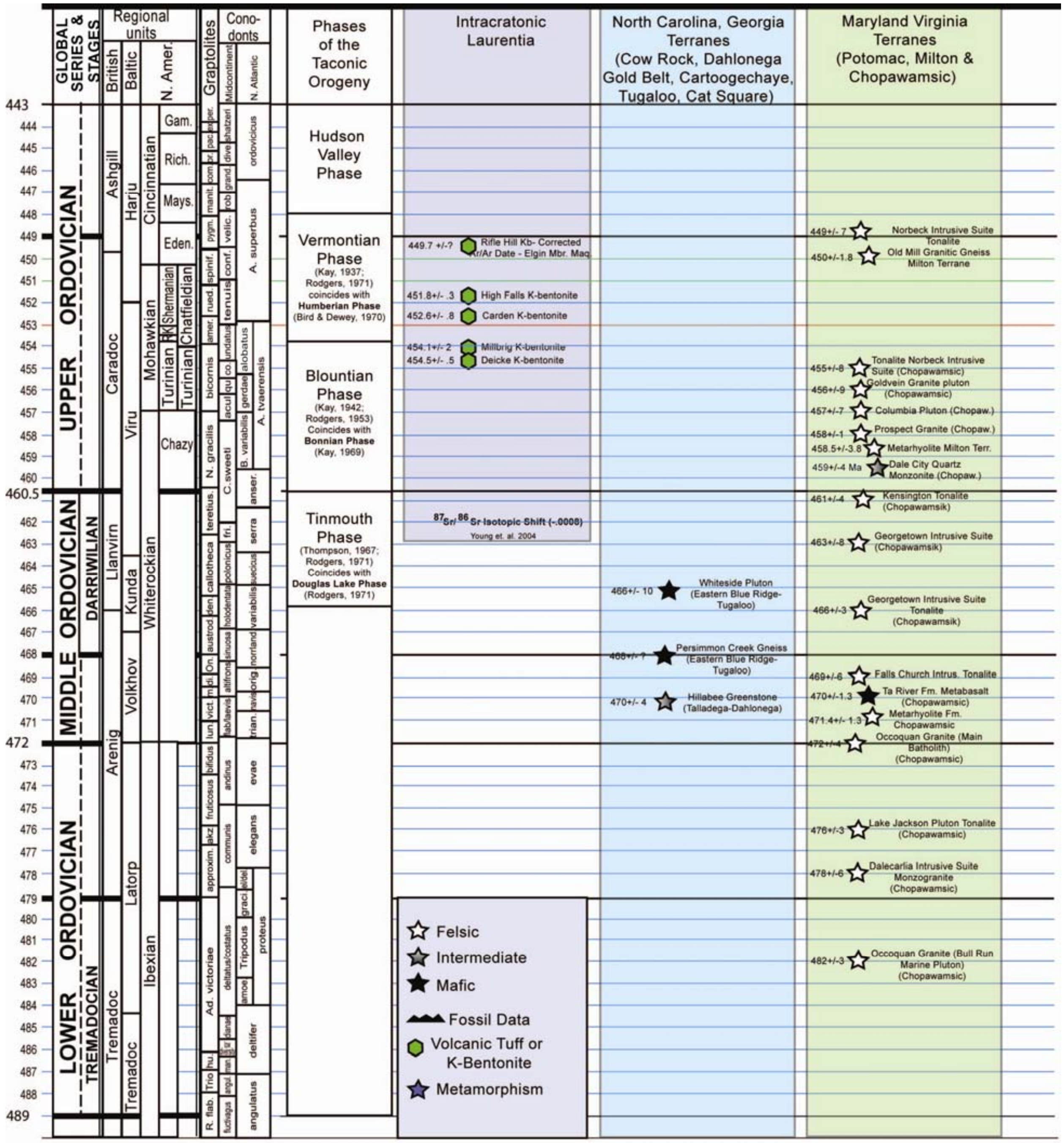


Figure 9: Chronology and absolute dates for volcanism and collision related events in terranes recognized in the central and southern Appalachians (mostly Blue Ridge and equivalents). Radiometric dates taken primarily from Coler and colleagues (2000) and references therein. As shown, Middle Ordovician volcanism in the southern Appalachians was dominated by intermediate to mafic events, whereas Early Ordovician to earliest Late Ordovician volcanism in the central Appalachians (Virginia-Maryland-Pennsylvania) was dominated by formation of felsic plutons, and intrusive suites therein. As with the Shelburne Falls Arc to the north, evidence for volcanism in these terranes ceased just prior to the end of the Blountian tectophase. During the Vermontian tectophase, after deposition of the Deicke and Millbrig K-bentonites, very little volcanism is recorded in these terranes.

Chatfieldian (Rocklandian-Kirkfieldian), and contributed to further thickening and eventual closing of the “marginal basin” (or forearc basin) that separated the accretionary wedge from the volcanic arc/micro-continent complex. Thus, in addition to evidence from relatively recent absolute ages of volcanic terranes in the Taconic Orogen as shown in figure 9, and the sedimentologic and structural evidence from the foredeep and accretionary wedge complex (as suggested by Lash and Drake), there is increasing evidence for subduction zone reversal coincident with deposition of the earliest Trenton Group. Most importantly, it can be inferred from these studies that the reversal event accommodated the continued development of the accretionary wedge via the closing of the “marginal” forearc basin by the Edenian. Given theoretical modeling of subsidence-inducing load geometries (see above), and data from provenance studies (discussed below), the much greater magnitude and duration of the complete basin filling episode of the later Vermontian tectophase (~10 million years as compared to the Blountian tectophase that filled in circa 5 million years) is likely attributed to the pronounced growth of the tectonic load as proposed for the first time herein. In this scenario, not considered by any previous model, the taper angle (height to width ratio) became substantially greater than during the Blountian event and upon narrowing, the load produced a much more substantial foreland basin than at any previous time. The foredeep, once depressed, subsequently migrated inboard with the accretionary wedge complex from its stalled location at the close of the Blountian.

Finally, a third problem with previous models for the Taconic Orogeny relates to its failure to make the connection between the Blountian and Vermontian tectophases and events recorded within craton interior regions. Thick successions of deep-water shales and flysch deposits incorporated into the accretionary wedge of the Martinsburg foreland, and eroded later in the Vermontian tectophase, were deposited much earlier and were in part age equivalent to sediments deposited in the Sevier basin further south. As noted by Shanmugam and Lash (1982), both of the Taconic foreland basin-forming tectophases (Blountian and Vermontian) appear to be analogous in their general stratigraphy. Nonetheless, the relationship between southern and northern phases has still not been established. Moreover, critical differences exist in the origin of sedimentary grains deposited in these two basins.

Phases of the Taconic Orogeny: analogous or not?

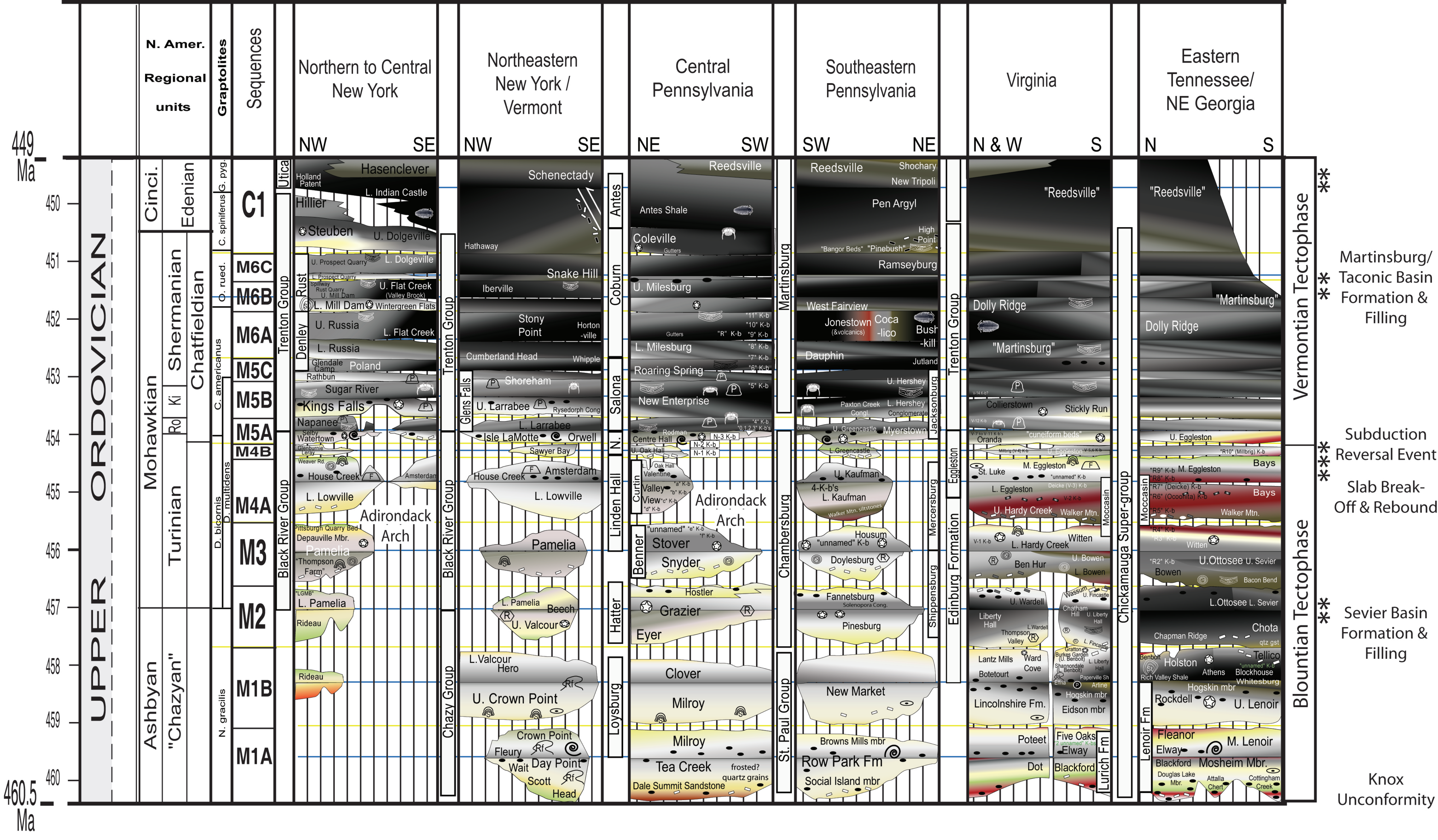
The decline of carbonate deposition and subsequent dark shale to flysch to red bed fill, in the southern Appalachians (Sevier Foreland Basin), was followed by the decline of carbonate deposition in the north-central Appalachian region (Martinsburg/Taconic Foreland Basin). These events were distinctively offset in time and have been considered representative of analogous collisional events. Based on the stratal packaging and sedimentary fill similarities, it was postulated that similar tectonic mechanisms were responsible for the development of both foredeep basins; albeit diachronously from Tennessee to Pennsylvania (Shanmugam & Lash, 1982). However, the Martinsburg-basin filling, from basin subsidence through flysch and molasse deposition, took significantly longer (>10 million years) and influenced a much greater area. In addition, only the last tectophase resulted in the final demise of the GACB.

An important aspect of the Shanmugam and Lash (1982) model was the occurrence of widespread unconformities overlain by deepening-upward carbonates in the base of each basin prior to shale and flysch deposition. Unconformities, following the studies of Jacobi (1981), Quinlan and Beaumont (1984), Ettensohn (1991), and Diecchio (1993), were attributed to the development of an uplifted peripheral bulge followed by regional beveling across the bulge to produce the unconformity –as discussed earlier. Hypothesized to have formed in response to "jamming" of the subduction zone during collision of the Laurentian margin, these unconformities were then subsequently buried during rapid subsidence as the foredeep basin and associated load migrated cratonward as the subduction zone became "un-jammed."

In the Shanmugam and Lash (1982) model, the basal Sevier Basin unconformity was recognized as the contact between the Lower Ordovician Knox Group and the overlying Upper Ordovician Lenoir Formation (Ashbyan), i.e. the Knox Unconformity (**figure 10**). In contrast, the basal Martinsburg Basin unconformity was recognized in eastern Pennsylvania, by the contact between the Jacksonburg Limestone (early Chatfieldian) and underlying Ashbyan strata along the Adirondack Arch. In this case, the unconformity represents much of the Turinian Stage as discussed above, but is of limited lateral extent compared to the earlier Knox unconformity. Even though both have been attributed to subduction "jamming" episodes in the Shanmugam and Lash model, the differences in scale of the unconformities is problematic.

It is clear that the Sevier Foreland Basin had not only formed prior to the pre-Jacksonburg unconformity, but it was much smaller spatially and was rapidly filled during the Turinian when the unconformity was formed along the Adirondack Arch in eastern Pennsylvania to southeastern New York. Moreover, as shown in the sequence stratigraphic analysis in Chapter 7, the Sevier Basin was overfilled with molasse well before major subsidence and clastic sedimentation commenced in the Martinsburg Basin to the north. Thus, although components of the uplifted accretionary wedge were certainly coeval with Sevier Basin sediments, the sedimentary fill within the Martinsburg Basin is distinctively younger than that of the Sevier Basin. Nevertheless, Asynchronous unconformities, the asynchronous sedimentary fills, and distant depocenters suggested to Shanmugam and Lash that the two basins were formed as two

Figure 10: Correlated sequence framework for GACB marginal areas nearest the Taconic Orogen showing basin formation and filling episodes for the Blountian and Vermontian tectophases (from Chapter 7). Key events in the orogen discussed herein are shown as are the positions of four strontium isotopic excursions as denoted by *.


N. Amer. Regional units
Graptolites
Sequences
Northern to Central New York
Northeastern New York / Vermont
Central Pennsylvania
Southeastern Pennsylvania
Virginia
Eastern Tennessee/ NE Georgia
449 Ma
460.5 Ma
UPPER ORDOVICIAN
Mohawkian
Ashbyan
"Chazyan"
Turinian
Shermanian
Chatfieldian
Cinci.
Edenian
C1
M6C
M6B
M6A
M5C
M5B
M5A
M4B
M4A
M3
M2
M1B
M1A
Trenton Group
Black River Group
Chazy Group
Adirondack Arch
Martinsburg
Chambersburg
St. Paul Group
Edinburg Formation
Chickamauga Super-group
Vermontian Tectophase
Blountian Tectophase
Martinsburg/ Taconic Basin Formation & Filling
Subduction Reversal Event
Slab Break-Off & Rebound
Sevier Basin Formation & Filling
Knox Unconformity

different, yet analogous, phases of the Taconic Orogeny.

More recent observations suggest that, despite similarity in sedimentary facies, depositional structures, and stratal packaging as documented by Shanmugam and Lash (1982), there is evidence for distinct dissimilarities between these basins that suggest “non-analogous” evolutionary histories for these basins. As alluded to above, it is difficult to explain the pronounced Knox Unconformity and associated deformation over wide areas compared to the relatively small-scale pre-Jacksonburg unconformity using the same subduction “jamming” scenario. Moreover, the timing of extensive volcanism (as recorded by volcanic ash beds) at the close of the Blountian and onset of Vermontian, as well as chemostratigraphic evidence (see below) suggest that both unconformities did not form from analogous subduction jamming episodes. In fact, subduction may have accelerated at the end of the Turinian rather than becoming completely “jammed” as suggested in earlier models. Additional evidence comes from quantitative and qualitative provenance studies of the Sevier and Taconic Basins (Bock et. al, 1998; Mack, 1985; Rowley & Kidd, 1981; Stephens & Wright, 1981). These studies provide a variety of observations that suggest seemingly subtle differences in sediment composition within these basins. Until now, these differences have not been considered in the context of models for the Taconic Orogeny and its tectophases. Herein it is suggested that these differences actually have pronounced implications for the evolution of the Taconic Orogeny.

Provenance Studies and their implication for evolution of the orogeny

Mack (1985) investigated the provenance of the sedimentary fill of the Blount clastic wedge in the Sevier Basin and compared his data to that of previous reports from the Martinsburg/Taconic basin. Sandstones from the Blount clastic wedge (Georgia/Tennessee) were derived primarily from quartz-dominated sandstones, micritic carbonates, illite-rich shales, and minor chert sources. Essentially this package of sediments was derived from older Cambro-Ordovician passive continental margin sources that were uplifted and eroded during the Blountian tectophase. A secondary suite of grains was derived from feldspar-rich sources indicating plutonic or high-grade metamorphic rock sources. Mack (1985) suggested that these may have come from basement sources, or uplifted Precambrian arkosic sands. Likewise Bock and colleagues (1998) suggested, on neodymium geochemical evidence, that these materials were likely derived from a Grenville Province source. A tertiary suite of grains apparently originated from low-grade metamorphic rocks as indicated by foliated metamorphic rock fragments (quartz and muscovite) and polycrystalline quartz. These sedimentary components were most likely derived from low-grade, weakly metamorphosed slates and phyllites. Collectively these data provide evidence that late Ashbyan to Turinian sediments of the Sevier Basin and its equivalents to the north were supplied primarily from passive margin continental shelf and slope deposits as they were uplifted during collision and formation of an accretionary prism. As such, many of the sediments from the Taconic allochthons, i.e. the Austin Glen Formation to the north, were also derived from similar source rocks, and potentially even the same accretionary prism during the same time (*N. gracilis* to *D. bicornis*-zones).

Samples from the Taconic/Martinsburg basin were also dominated by lithic fragments comparable to those in the Sevier Basin. However, Mack (1985) suggested that sandstones in the Taconic/Martinsburg basin were enriched, relative to the Blount, in grains derived from intermediate to mafic volcanic sources. This variation in clastic wedge composition was considered to have been the result of: 1) variations in tectonic style along strike, 2) variations in the distribution of rock types within source terrains, or 3) differences in the level of erosion between similar orogenic terrains (Mack, 1985). Although not considered by Mack, in light of growing evidence for a change in subduction direction as discussed, a fourth explanation might be related to telescoping of the accretionary wedge-volcanic arc during the same but protracted collisional event. This would have increased the supply of volcanic source rocks to the foreland basin – especially if the "marginal" forearc basin became closed as surmised by Lash and Drake (1984). Of course wind-blown volcanic ashes were deposited in the Sevier Basin, but sandstones show that mafic-derived sediments were not supplied in major quantities during the Blountian tectophase.

Another study (Andersen, 1995) investigated the provenance of mudstones from both the Taconic and Sevier foreland basins using whole-rock geochemistry and clay mineral composition. Andersen noted that there is an upward increase in the proportion of mafic source rocks into the Taconic foreland basin. Mirroring the study by Mack (1985), sedimentary signatures of mudstones suggest that the Sevier foreland contained "comparatively lower concentrations of mafic elements" than the Taconic foreland. This observation is consistent with sediment sourcing initially from a non-magmatic source terrane followed by sediment sourcing from a slightly more magmatic terrane.

Taken collectively, data from Mack (1985) and Anderson (1995), show that it is increasingly likely that Blount sediments were derived primarily from an emergent accretionary wedge complex formed during subduction of the distal Laurentian continental margin. In turn, younger Vermontian tectophase sediments, although still reflective of significant accretionary wedge sources, show a greater influence of magmatic arc signatures, consistent with previous suppositions. Given the age of key volcanic terrains and inferred paleoceanographic circulation patterns in this region (Wilde, 1991), a forearc basin likely formed between the accretionary prism and magmatic arc (i.e. see figure 2). The forearc basin would have kept sediments derived from the volcanic arc segregated from the adjacent foreland basin. Thus, until the forearc basin was filled or closed, sediments in the foreland basin would not reflect a magmatic source, except for the occurrence of wind-blow volcanic ash, and would remain distinct (i.e. Blountian tectophase). In contrast, once the forearc was filled with sediments or closed by tectonism (i.e. Vermontian tectophase as discussed herein), magmatic-influenced sedimentation could spill over the accretionary wedge and enter the foreland basin.

CHEMOSTRATIGRAPHIC EVENTS & THEIR IMPLICATION FOR RATES OF SEA-FLOOR SPREADING AND ACCRETIONARY WEDGE DEVELOPMENT

As documented in detail in Chapter 6, important chemostratigraphic events are now being recognized across the GACB and have been refined relative to major chronostratigraphic intervals of the platform and foreland basin. A number of important isotopic excursions (strontium, neodymium, and carbon) are now recognized to record significant changes in sedimentary provenance and chemical cycling of Ordovician oceans and marginal seas. Two primary isotopic systems, $^{87}Sr/^{86}Sr$ ratios (see Chapter 6, figure 18), and ε_{Nd} values (see Chapter 6, figure 20) observed in the GACB have pronounced implications for the correlation of sedimentologic changes occurring within the Taconic foreland basin system with events occurring in the Taconic hinterland. Moreover, it is suggested herein, that these isotopic excursions are tied to major pulses in tectonism of the Laurentian margin as it underwent plate tectonic reorganization during the Ashbyan to Cincinnatian.

$^{87}Sr/^{86}Sr$ Excursions

As discussed in chapter 6, there are four paired $^{87}Sr/^{86}Sr$ excursions tentatively-recognized from strata in the CBRT interval (**see figure 10**). Each event shows a pronounced drop in $^{87}Sr/^{86}Sr$ values followed by a pronounced increase in $^{87}Sr/^{86}Sr$ values. Rapidly declining values are thought to reflect increased volumes of strontium input to the global ocean from hydrothermal leaching of basalts. This is most likely attributable to increased rates of sea-floor spreading. In contrast, increasing $^{87}Sr/^{86}Sr$ values, are thought to reflect increased rates in the delivery of strontium weathered from continental crust and related sources. During the approximately10 million year interval encompassed in this study, three prominent excursions are noted, along with a smaller, less prominent excursion.

The first occurs after the Knox Unconformity and is roughly coincident with the onset of major deepening in the Blountian tectophase near the Ashbyan-Turinian boundary. A surge in sea-floor spreading at this time may have contributed not only to higher global sea-levels, but also may have contributed to the rapid building of the accretionary prism and flexure of the Laurentian margin. This led to subsidence of the Sevier basin and isolation of craton-interior areas inboard of the forebulge where carbonate sedimentation persisted. This rapid growth of the accretionary prism, after it became emergent, enabled significant volumes of old, exhumed passive margin sediment to be remobilized and shed into the Sevier basin and its equivalent to the north during the later Ashbyan to Turinian interval. The re-introduction of these sediments would account for the rapid increase in $^{87}Sr/^{86}Sr$ values immediately following the surge in seafloor spreading.

The second pronounced excursion ranges across the Turinian-Chatfieldian boundary. This excursion first shows a lowering of $^{87}Sr/^{86}Sr$ values leading out of the Turinian and into the earliest Chatfieldian during the transition into the Vermontian tectophase (**see figure 10**). In this case, it is apparent that a major pulse in seafloor spreading introduced large volumes of basalt-derived strontium into the Ordovician oceans at the end of the Turinian coincident with eruption and deposition of the Millbrig K-bentonite. If, as suspected by Karabinos and colleagues (1998), this interval is suspected to reflect subduction reversal, and despite sedimentologic evidence of eustatic lowering, the occurrence of this rapid increase in sea-floor spreading may indicate decoupling (slab break-off) of the east-dipping slab and initiation of west-dipping subduction further east. The overall shallowing and increasingly uniform conditions across much of the eastern GACB leading into the latest Turinian is inferred here to record rebound of the cratonic margin-wedge complex (as thought to be occurring in Timor region today) at the close of the Blountian tectophase after the east-dipping slab detached suddenly. This is distinctly different than any previous model, including that of Shanmugam and Lash (1982) that predicted a "jamming" of the subduction zone at this time. Moreover, this rebound would account for the rapid change from flysch to molasse in the Sevier Basin (identical to a model proposed by Sinclair, 1997 for the type-Molasse succession within the North Alpine foreland basin of Europe). It would also account for rapid progradation of molasse facies into carbonate-dominated facies along the northwestern edge of the Sevier Basin. Furthermore, rapid melting of the detached, east-dipping slab may have provided the rapid volume of melt necessary for the eruption of the Hagan K-bentonite complex which includes the Millbrig K-bentonite and its possible equivalent the Kinnekulle of the Baltic region.

Following the late Turinian-early Chatfieldian drop in $^{87}Sr/^{86}Sr$ values, they immediately rebounded for the second half of the paired excursion, reflecting another increase in continent-derived strontium at a time of significantly higher sea-levels. This isotopic change is coincident with: 1) the first widespread-deepening in the Trenton observed across the GACB and 2) the introduction of siliciclastics at the outset of the Vermontian tectophase. Thus, it is proposed herein that renewed thrusting in the hinterland resulted in an amplification of the tectonic load (accretionary prism) at the continental margin, and this led to another phase of subsidence. In this case, in contrast to the Blountian tectophase, the much more substantial growth of the accretionary prism may have been triggered by the closure of the forearc basin, as stacked-thrust sheets overrode the accretionary wedge formed in the earlier tectophase.

The third paired excursion, the weakest of the four (but with nearly the lowest overall $^{87}Sr/^{86}Sr$ values of this part of the Ordovician), is noted in the mid to late Chatfieldian (late *C. americanus* to *O. ruedemani* zone), and approximately coincident with the M6A sequence as discussed elsewhere. The lowering of $^{87}Sr/^{86}Sr$ values, follows a major shallowing at the end of the M5 sequence, and is coincident with a major deepening recorded in GACB strata. This again suggests another short-term surge in sea-floor spreading following a period of lower sea-levels. Interestingly, this surge is timed with the occurrence of soft-sediment deformation "seismites" across much of the platform as well as in foreland basin settings. Evidently, increased rates of seafloor spreading, following a period of quiescence at the end of the M5 sequence, not only contributed to higher sea-levels, but also produced movement on numerous faults both within the hinterland as well as in proximal areas of the foreland basin (i.e. Jessamine Dome). This compressional episode renewed subsidence in some areas of the foredeep basin and intracratonic troughs, i.e. the Sebree Trough.

$^{87}Sr/^{86}Sr$ isotopic values rebound again immediately following the negative pulse indicating a renewed pulse of siliciclastic-derived strontium was introduced back into the oceans (**see figure 10**). In many areas, including in the foredeep and on the periphery of the foreland basin, the increased progradation of slightly coarser-grained siliciclastics has been noted, i.e. Snake Hill of eastern New York, Ramseyburg of eastern Pennsylvania, Millersburg of Kentucky, etc. and is now timed to this excursion. It is inferred here that the previous increase in sea-floor spreading led to additional thrusting and uplift in the hinterland which enabled an increased rate of weathering and consequently a greater supply of siliciclastics.

The final paired $^{87}Sr/^{86}Sr$ isotopic event occurred in the Edenian (latest *C. spiniferus – G. pygmaeus* graptolite zone). This event appears to be coincident with: 1) another major pulse of sea-floor spreading, 2) higher sea-levels (likely the highest of the Ordovician), 3) a renewed pulse in deformation, and is again followed by a longer-term change back to continent-derived strontium. In this case, the pronounced return to siliciclastic progradation is coincident with deposition of the upper Kope Formation including the Garrard Siltstone of Kentucky, the Schenectady Formation of New York, the Reedsville Formation of Pennsylvania, etc. This final event marks the approximate end of the Vermontian tectophase and the onset of the Hudson Valley phase that records the flysch to molasse filling phase of the later event (see Chapter 1, figure 5). Thus beginning in the Edenian, the remainder of upper Ordovician strata record the progradation of the massive Queenston Delta and infilling of the Martinsburg Foreland Basin through the close the Hudson Valley phase.

A NEW INTEGRATED MODEL FOR THE EVOLUTION OF TWO FOREDEEP BASINS DURING THE TACONIC OROGENY IN RESPONSE TO SUBDUCTION ZONE REVERSAL

Constraints on the Timing of development of the Taconic Foreland Basin

Changes in GACB Platform: Rimmed to Ramped Morphologies

As discussed, the accentuation of the Knox Unconformity locally and regionally across the GACB is interpreted to represent up-warping of the crust owing to tectonic collision as a result of seizure on the subduction zone (Jacobi, 1981; Shanmugam & Lash, 1982, Ettensohn, 1991, Finney et al., 1996). Cady (1945) and Fisher (1954) presented evidence suggesting that broad folding, normal faulting, and possible wrench faulting, associated with the westward migration of the west Iapetan trench, may have been responsible for the increased scale of the unconformity in some regions especially in New York to Vermont (Fettke, 1948). Despite assertions that the unconformity was entirely tectonic in construct, it is recognized in many regions across eastern and western Laurentia (Sloss, 1963; Mussman and Read; 1986; Ross and Ross, 1992), and even globally (Hallam, 1992; Nielsen, 2004). Thus, this unconformity records evidence for both tectonic modification of the platform and a eustatic sea-level drop associated with slowing of sea-floor spreading rates at the end of the Pan-African Orogeny (Shields & Veizer, 2004).

Read (1989) suggested that there is sedimentologic evidence in the eastern GACB that late Beekmantown (pre-Knox Unconformity) deposition saw the simultaneous development of prograded ramp morphologies along the eastern margin of North America in the southern, central and northern Appalachians. These ramp morphologies were in strict contrast to earlier rimmed platform morphologies of the same region. Read argued that the lowering of sea-level during the transition to the end of the middle Ordovician contributed to progradation of facies and ramp formation. Yet, he also noted rapid shallowing in distal slope regions that could not be explained by depositional filling and proposed that ancestral listric normal faults (which had been subsiding) began to reverse in off-shelf, deep water slope areas. Thus, Read argued that prior to eustatic sea-level fall at the termination of the Sauk Megasequence, there is evidence for off-shelf shallowing, and cessation of continent margin subsidence signifying initial phases of collision of the Taconic Orogeny, and the initial uplift of the peripheral bulge.

Recent biostratigraphic, sedimentologic, and sequence stratigraphic studies (Landing, 2000, Landing et al., 2003b) have also provided constraints for these early developments in the Taconic Orogeny from the New York Promontory region. Landing and colleagues recognized patterns in deep-water slope and rise deposits off the eastern margin of the GACB that were attributed to both eustatic and tectonic events during the initiation of tectonic collision. Although later incorporated into the Taconic allochthons, faunal changes and color signatures of slope and rise deposits of Upper Cambrian to Middle Ordovician rocks can be attributed to sea-level change. Deep-water slope and rise sedimentation produced variations in deepwater mudstones ranging from condensed, pyrite-rich dark shales (sea-level highstands) to gray and green, radiolarian-rich mudstones (sea-level lowstands). As indicated in these studies, deposition of Deepkill black shales on the slope and rise was coincident with sea-level maxima of the pre-Knox Beekmantown highstand (early Darriwilian) and roughly timed with the change from rimmed platform morphologies to ramp morphologies as noted by Read (1989). Subsequently, deposition of green mudstones (lower Mt. Merino Formation) commenced in the late Darriwilian (late *teretiusculus*- early *gracilis* graptolite zones) coincident with the Knox lowstand and exposure. Immediately following the lowstand, yet another sea-level rise in the Ashbyan (late *N. gracilis* zone) is documented in the shift from the lower green Mt. Merino to the upper black Mt. Merino shales. This is coincident with deposition of the Chazy Group and the formation of the Sevier Basin in the south.

As mentioned, the Indian River red mudstones were previously inferred to represent the "terra-rosa" lowstand detritus derived from erosion of the Knox Unconformity (i.e. as proposed by Bird & Dewey, 1970), or post-Chazy lowstand detritus (i.e. as proposed by Rowley & Kidd, 1981). Although poorly dated themselves, these red mudstones are located stratigraphically above the black Deepkill shales (pre-Knox Highstand) and below the lower "green" Mt. Merino Formation (Knox Lowstand). Thus, the Indian River succession is clearly pre-late *teretiusculus* Zone and therefore was deposited prior to the Knox unconformity and therefore the succession does not represent the "terra-rosa" of earlier models. Enigmatically, the Indian River red mudstones were deposited during the pre-Knox highstand. Landing and colleagues proposed that Deepkill and similar black shales were developed in deepwater, dysoxic to anoxic environments under oxygen minimum conditions during highstands. Obviously, the red-mudstones were not deposited under such reducing conditions. Therefore, Landing (2003, pers. com.) has surmised that the Indian River was deposited on an uplifted portion of the seafloor (i.e. cratonward migrating forebulge), that rose above the oxygen minimum zone and resulted in the accumulation of starved, iron-rich mudstones, radiolarites, and even very early K-bentonites.

These observations, combined with the data from Read (1989), indicate that the initial development of the forebulge occurred in deep-water environments off the coast of eastern Laurentia sometime prior to the Knox lowstand. Subsequent migration of the Indian River forebulge up the continental slope and rise was manifest as the reversal of ancestral listric faults outboard of the shelf margin immediately prior to the Knox Unconformity resulting in a change in carbonate platform morphologies. The peripheral bulge then migrated onto the continental shelf coincident with the Knox exposure. Given the width of the foredeep basin (~150 km for the modern Timor-Banda orogen), it appears that the trench and the leading edge of the accretionary wedge itself would have likely initiated its collision with the continental slope and rise at this time. If crust of the distal continental slope was unable to move into the subduction zone due to its buoyancy, these materials and their mantle of Cambro-Ordovician rocks could have indeed "jammed" or at least slowed movement along the subduction zone as postulated in earlier models. Moreover, a subduction failure may have attributed to a much more intense doming of the peripheral bulge and expanded the spatial scale of the Knox Unconformity. Clearly this did not happen in the same fashion during the Vermontian, as the pre-Jacksonburg unconformity was not as significant.

As the accretionary prism moved up the continental slope in the Ashbyan, and rapidly so after the Ashbyan-Turinian strontium shift, the incorporation of slope sediments into the base of the accretionary prism would have contributed to the rapid growth of the prism at the outset of the Blountian tectophase. Furthermore, the growth of the accretionary wedge not only provided the load to induce the formation of the Sevier Basin, but the wedge may have become emergent for the first time thus providing a major source of cannibalized sediment derived from the once passive margin. These sediments thus produced the first flysch deposits in the Sevier Basin, and also on the slope and rise further north (i.e. Austin Glen Greywacke).

Collision and foreland basin migration trajectory.

Most paleogeographic models for the Early to Middle Ordovician, based primarily from those of Scotese and McKerrow (1991), show a single elongate, east-dipping subduction zone along the entire southeastern margin of Laurentia (**figure 11**). Immediately to its west lie the Taconic Foreland Basin, and the GACB. To the east of the subduction zone lays an elongate microcontinent/volcanic arc complex. Despite this simplistic reconstruction, the geology of the Appalachians shows a much more complex scenario with some evidence for along strike variation in the orogen.

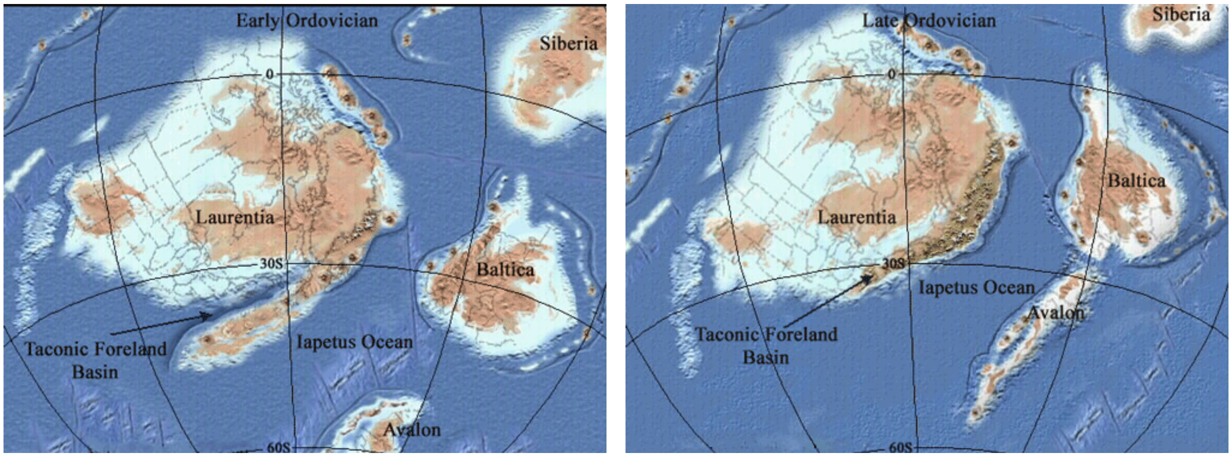


Figure 11: Paleogeographic maps for the Early (left) and Late Ordovician (right) modified after Blakey, (2007). The position of the Taconic Foreland Basin, and the east-dipping subduction zone on its eastern margin, is shown along the entire southeastern edge of Laurentia during the Early Ordovician. Equatorward, Blakey depicts a west-dipping subduction zone off the easternmost margin of Laurentia in the location of the Canadian Maritimes. Outboard (southeast) of the foreland, an elongate microcontinent/volcanic arc complex is also depicted. Another east-dipping subduction zone is shown on the eastern edge of the Iapetus Ocean (western margin of Baltica). By the Late Ordovician (circa 450 ma), the microcontinent/ arc complex is completely sutured to the GACB. Although, the west-dipping subduction zone is still shown further north in the location of the Canadian Maritimes, no further plate tectonic boundary is depicted in the western Iapetus. The west Baltic subduction zone has extended southward and merged with the west Avalon subduction zone.

As suggested by Thomas and colleagues (2001), there is very little evidence in the south-central Appalachians (North Carolina to Tennessee) for a volcanic arc complex like those recognized farther north (i.e. Chopawamsic, Potomac, West Minster, Baltimore, Manhattan Prong, Shelburne Falls, etc.; **see Chapter 1, figure 11**). To the south on the Alabama Promontory (Georgia to Alabama), thrust sheets of the Blue Ridge-Piedmont Province (Dahlonega Gold Belt, Hillabee Greenstone, Persimmon Creek Gneiss, etc.) do show evidence for mafic, to intermediate volcanic rocks (**see figure 9**) indicating the presence of a volcanic terrane in this region, although primarily of mafic composition. Volcanic terranes (primarily intermediate to felsic) once again appear in the Blue Ridge – Piedmont along the Virginia Promontory, and along the southeastern margin of the New York Promontory and farther north (**figure 12**). Given this arrangement, it appears that many of the volcanic terranes appear to be located on and near southeastern facing promontories, while the North Carolina-Tennessee region (i.e. Tennessee Embayment) seems to be a window within which no record of volcanism

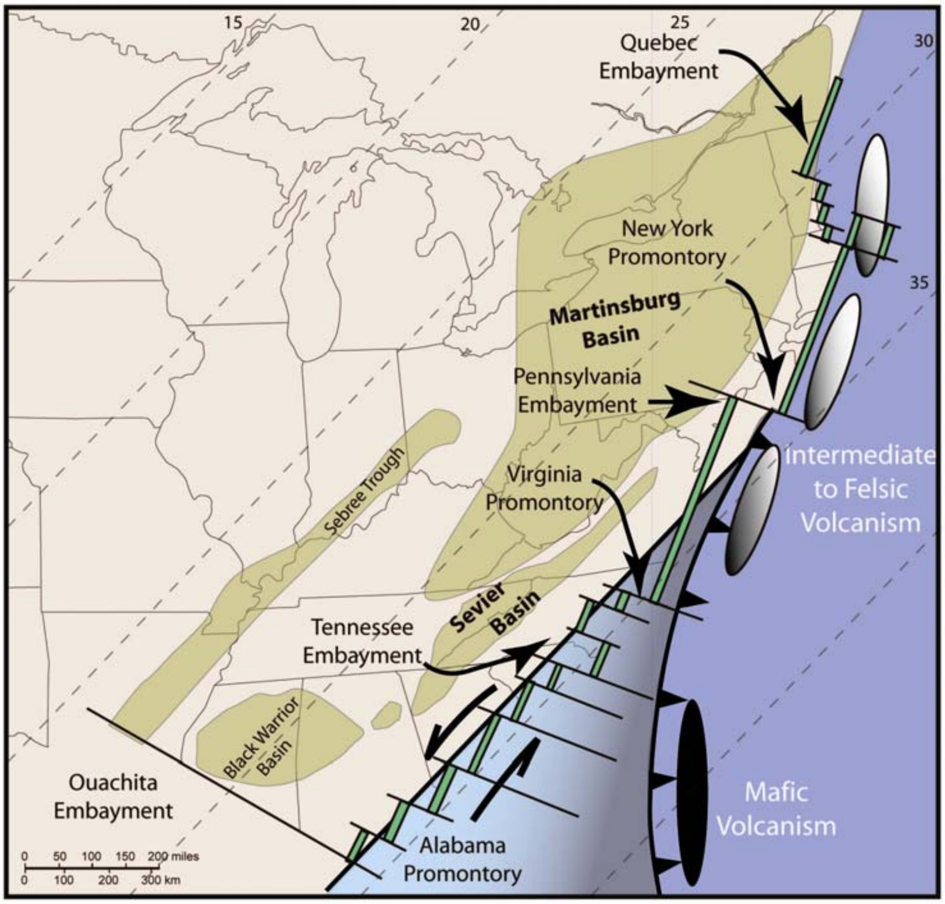


Figure 12: Plate tectonic configuration for the eastern margin of Laurentia in Middle Ordovician time, prior to the Taconic Orogeny. Laurentian continental plate rift margin is shown in green as derived from Gao et al., 2000. The most recent plate tectonic reconstructions for this portion of the Ordovician (i.e. Scotese, 1997) and drawn by Ettensohn et al., 2002b, show a small wedge of oceanic crust sandwiched between the Laurentian plate and the Iapetus Ocean plate. This small wedge is interpreted to subduct toward the northeast beneath the Iapetus Ocean plate east producing predominantly mafic volcanism. The western margin of the wedge is inferred to be dominated by left-lateral transform movement along the southeastern margin of the Laurentian Plate. Farther north, volcanism is dominated by intermediate to felsic volcanism.

is recorded, sedimentologically or otherwise (Mack, 1985; Thomas et al., 2001). Thus it appears that significant, along-strike variation did exist along the length of the Taconic Orogen, with a more mafic southern prong, and a more felsic northern prong. These observations have suggested the presence of a small oceanic wedge between the Iapetan Ocean plate and the southeastern margin of the Laurentian plate (Scotese, 1997; Ettensohn et al., 2002b). Ettensohn and colleagues (2002b) show a north-south oriented subduction zone with northeastward subduction on the eastern margin of the oceanic wedge plate. This northeastward subduction is accommodated by the presence of left lateral transform movement along the western margin of the oceanic wedge and the southeastern margin of the Laurentian plate. Although this configuration would explain mafic volcanism to the south, it still falls short of explaining the initiation of the Blountian tectophase and the formation of the Sevier Basin prior to the northern tectophase. In fact, this scenario and plate tectonic configuration suggests collision would have initiated in the vicinity of the New York Promontory prior to collision in the south, a scenario that is clearly not accurate.

Given the presence of Ordovician aged: 1) NE-SW (modern orientations) trending volcanic terranes adjacent to and near prominent southeast-facing promontories along the Laurentian plate margin, 2) numerous joint and fault sets with NE-SW trends, and 3) a number of NW-SW lineaments, joints, and a few faults oriented perpendicular to the former sets (i.e. see Chapter 1, figure 13), it appears that the trajectory of collision, especially in the southern to central Appalachians, was dominantly northwestward rather than north-south and may have been oblique (**figure 13**). This trajectory may have evolved slightly to a more north-south orientation after the initial phase of collision in the south due primarily to the irregular continental margin. The trajectory of collision, and incidence angle of forebulge movement, must be constrained relative to areas showing simultaneous accentuation of the Knox exposure. As such, relatively extensive uplift and truncation of pre-Knox strata along the Laurentian margin has been primarily reported in the vicinity of promontory regions (i.e. Alabama-Georgia, Virginia, and New York). Whereas protected reentrant regions, such as the Tennessee and Pennsylvania embayments, show evidence for more continuous deposition and a significantly shorter duration for the Knox Unconformity. Thus, it is suggested that the collisional front and the strike of the subduction zone may have been oriented as shown in figure 13. Moreover, given this orientation it is still possible to produce mafic volcanism in the south from subduction of the leading edge of

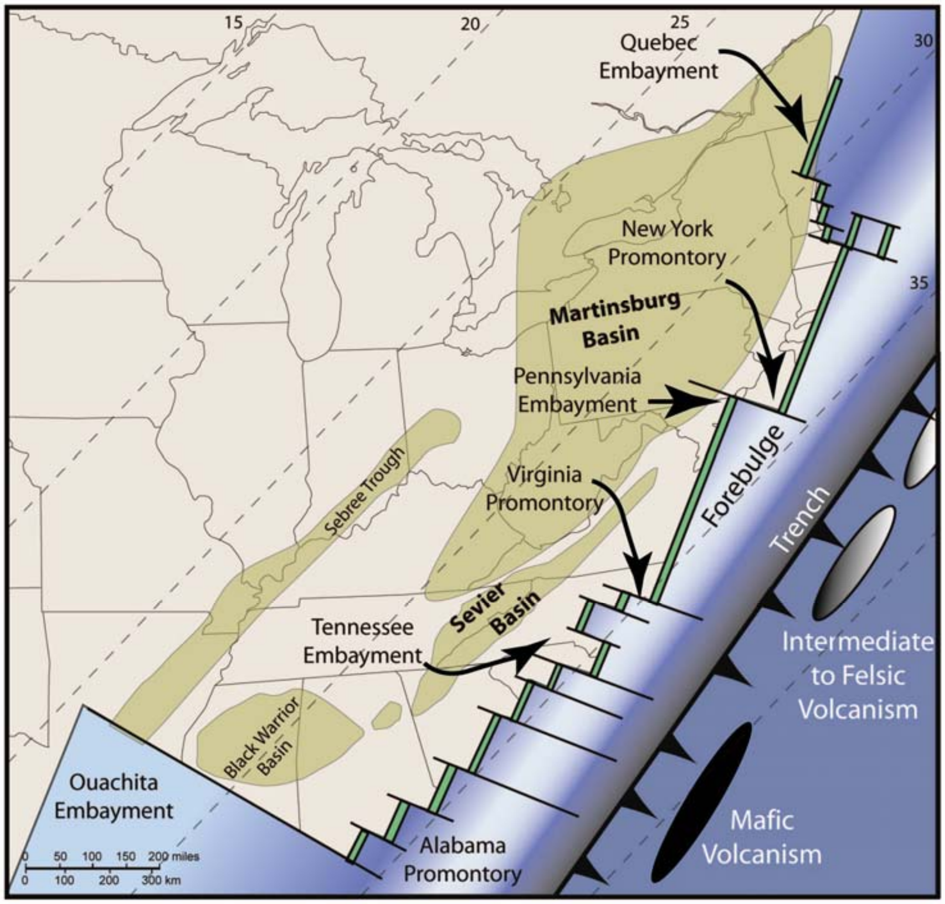


Figure 13: Plate tectonic configuration for the eastern margin of Laurentia in Middle Ordovician time, prior to the Taconic Orogeny to be compared with figure 12. Laurentian continental plate rift margin is shown in green as derived from Gao et al., 2000. The position of the trench (pre-Taconic Foreland Basin) and associated forebulge are shown as they approach the southeastern edge of the Laurentian craton immediately prior to the Knox Unconformity. Southeast of the trench, numerous volcanic terranes have already formed and/or are in the process of forming. Given this trajectory, the forebulge makes contact with the larger promontories first and would explain reversal along ancestral listric faults and the change from rimmed carbonate platform to ramp morphologies, as noted by Read (1989).

oceanic crust attached to the Laurentian plate margin. Further north, it is plausible, as suggested by Lash and Drake (1984), that a small microcontinent (i.e. a failed rift block) of felsic composition may have been caught in the subduction zone thus providing for felsic to intermediate volcanism in that region.

Constraints from theoretical models and the modern Timor-Banda Orogen

Given the pulsed nature of the Taconic Orogeny, with its less invasive southern tectophase, and its more substantial and pervasive northern tectophase, a number of theoretical and empirical observations governing the scale and dimensions of foreland basins provides for the following observation – implication scenarios that help further constrain the development of an updated, chronology for the Taconic Orogeny:

- Observation 1: Compositional properties of the Laurentian lithosphere, between the southern and northern margin of Laurentia, were probably not substantially different and are assumed to be of approximately the same age (Rodgers, 1971). Nonetheless, south of the Virginia Promontory, in the vicinity of the Tennessee Embayment, basement rocks appear to be substantially more fully dissected with narrower rift blocks and transform boundaries compared to the much broader Virginia and New York Promontories (Thomas, 1991)
- Implication: Given these structural differences, it is presumed that flexure of the foreland in the southern phase may have been accommodated through the reactivation and subsidence of closely-spaced, failed Iapetan rift segments allowing for a narrower foredeep. Farther north, more widely spaced rift segments may have favored the formation of a much broader foreland basin encompassing a greater area.
- Observation 2: Although the absolute depth of the Sevier Basin is not well constrained, carbonate interbeds and bioturbated calcareous mudstones are more prevalent throughout the basin fill, compared to the Martinsburg Basin, and apparently originate from westerly and northerly sources. This also suggests that the basin may have been narrower, and perhaps less deep than the later Martinsburg Basin.
- Implication 2: If the Sevier Basin was indeed shallower, theoretical modeling suggests that the taper angle (height to width ratio) of the load was also shallower (Garfunkel & Greiling, 2002). This would indicate that the leading edge of the load (accretionary prism) was not stacked very high.
- Observation 3: Provenance studies (Mack, 1985; Andersen, 1995; Bock et al., 1998) of Ashbyan-Turinian aged strata from the Sevier Basin and the Taconic allochthons indicate recycled sedimentary signatures, with relatively low metamorphic and mafic source signatures.
- Implication 3: It is likely that the load inducing the formation of the Sevier Basin was related to the formation of an accretionary prism that had, as yet, undergone very minimal metamorphism and compressional shortening.
- Observation 4: Geochemical, provenance, and paleocurrent analyses of sandstones and shales from the Austin Glen Member of the Normanskill Formation (Ashbyan to Turinian-aged "flysch") indicate that these

southerly-derived sediments were also dominated by recycled, sedimentary components, with little to no evidence for ophiolite, mafic volcanics, or high-grade metamorphics (Bock et al., 1998).

- Implication 4: A time-equivalent (of the Sevier Basin) foredeep was receiving flysch some distance outboard of the shelf-slope break southeast of the New York Promontory. It too is interpreted to have derived sediments from a relatively young accretionary prism that was further outboard of the cratonic margin. Based on the south-southwestern source of these sediments (Krueger, 1963), the Austin Glen flysch was likely delivered into the same coeval, but laterally extensive foredeep and were derived from the same accretionary prism that produced the Sevier Basin fill.
- Observation 5: By early late Turinian, the southern "Sevier-portion" of the foredeep had become overfilled to red beds (Bays/Moccasin Formations) and received extremely large volumes of volcanic ash. At this time, the time-equivalent foredeep in the New York Promontory region appears to have become shallower, although it was still receiving flysch as recorded by the uppermost Austin Glen. The interval also contains some of the lowest olistostromes (lithified blocks of carbonates) within the Taconic melange (Bosworth & Kidd, 1985). These appear to predate the deposition of the early Chatfieldian Rysedorph Hill conglomerates and breccias of New York, and the time equivalent Hershey conglomerates of Pennsylvania. Carbonate deposition continued on the New York Promontory, although they started to show signs of increased starvation.
- Implication 5: The Vermontian tectophase had not yet initiated major subsidence to form the Martinsburg Basin, although development of large southeasterly-derived olistostromes (from the accretionary prism) in the foredeep basin imply destabilization and over-steepening (via thrusting) in the accretionary prism and therefore an increase in the taper angle of the load. This coincides with the pulse of sea-floor spreading at the close of the Turinian and the voluminous volcanic eruptions discussed earlier – possibly the result of subduction reversal.
- Observation 6: Provenance studies from the Martinsburg Basin in the early Chatfieldian indicate the introduction of an additional source of siliciclastic sediments (i.e. from higher grade metamorphics, and intermediate to mafic volcanic sources; Hiscott, 1978; Mack, 1985; Andersen, 1995). Moreover, these siliciclastics are delivered across the GACB platform well up into the Upper Mississippi Valley during the M5A sequence highstand as shown by neodymium chemostratigraphy (Fanton & Holmden, 2007).

- Implication 6: It appears that the continental shelf in the northern Virginia to New York Promontory region was over-thrust by the accretionary prism inducing subsidence proximal to the load. Moreover, with the first orogen-derived sediments being introduced to the Martinsburg Basin, it is apparent that the accretionary prism may itself have been over-thrust, with successive sheets of higher-grade metamorphic and volcanic-influenced rocks derived from the forearc basin.
- Observation 7: The first "seismites" in the GACB interior, along with the formation of fault-bounded, craton interior features (i.e. the Sebree Trough), occur in the middle early Chatfieldian (Rocklandian). Pronounced destabilization in the New York Promontory via topographic inversion along ancestral fault blocks (i.e. Adirondack and Canajoharie Arches, and related block-faulted features), produced numerous deep-water intraclastic limestone deposits (i.e. Hershey, Rysedorph Hill, etc.), and synsedimentary deformation.
- Implication 7: These provide physical evidence for the migration of the structural front of the foreland basin (back bulge-forebulge complex?) some 300-400 km inboard of the cratonic margin and the rapid subsidence in the leading edge of the foredeep basin as it migrated inboard of the Turinian continental shelf-slope break.
- Observation 8: The first major, relatively un-interrupted succession of organic-rich dark, black shales in the Martinsburg foreland basin occur roughly coincident with the end of the *Amorphognathus tvaerensis* conodont biozone (early mid-Chatfieldian) as recorded by deposition of the Flat Creek Shales of New York, the Bushkill black shales of Pennsylvania, and the medial Martinsburg of West Virginia. These are correlated with condensed, dark-shales and allodapic limestones in more proximal foreland basin settings (i.e. the Brannon Member of the Lexington Formation in Kentucky and its equivalents) and are coincident with additional seismite occurrences.
- Implication 8: This deepening, is correlated with the third pulse in sea-floor spreading rates described from strontium isotopic data, and likely record both sea-level highstand and a renewed thrusting in the accretionary wedge-volcanic arc complex and subsequent subsidence in the foreland basin.
- Observation 9: The most widespread, and substantial deepening event in the Late Ordovician is recorded in the Edenian (*Geniculograptus pygmaeus* zone) by the deposition of the Indian Castle and Collingwood shales of New York-Ontario, the Antes and Pen Argyl Shales of Pennsylvania, and their correlatives in more proximal areas (i.e. the Kope Formation of Kentucky-Ohio). As before, numerous seismite horizons in the proximal foredeep (New York and Pennsylvania), and shallower regions of the cratonic foreland (i.e.

Kentucky and Ohio), are timed with these events as is the fourth pulse in sea-floor spreading. Also noted is a prominent discontinuity and sediment starvation surface in the foreland basin recording the deepening.

- Implication 9: This provides evidence for a much more vigorous deepening over a much broader area of eastern Laurentia than in any previous event. Based on theoretical modeling (Garfunkel and Greiling; 2002) the substantially larger scale of this pulse suggests that the foredeep may have become wider than in previous pulses. It is suggested that this deepening is not only eustatic, but the extremely condensed and dysoxic shale facies indicate that the basin was also deeper as a function of a much narrower load with a higher taper angle within the accretionary wedge -forearc-arc complex. This event may record the final phase of shortening.
- Observation 10: The occurrence of 1) Chatfieldian-aged volcanism outboard of the Shelburne Falls Arc east of the buried passive margin (Karabinos et al., 1998), and 2) the development of through-slab extrusive basaltic volcanics in the foredeep (i.e. Stark's Knob of New York; Landing, et al., 2003; and within the Hamburg Klippe in the vicinity of the Reading Prong of southeastern Pennsylvania, i.e. Jonestown Volcanics; Thompson, 1999) records a very substantial change in the architecture of the subduction zone during and after the Chatfieldian.
- Implication 10: At the very least, if the proposed reversal via slab break-off is accurate, the pattern of volcanism and the supply of magma would have been significantly altered by: 1) melting of the remaining components of the subducted ancestral Laurentian plate, and 2) newly initiated melting of the western edge of the newly subducting, west-dipping slab of the Iapetus Ocean plate. If subduction was shallow, this would have allowed the formation of both intermediate magmas near the subduction zone (i.e. Bronson Hill Arc) as well as mafic magmas further to the west (i.e. Starks Knob-Jonestown through-slab volcanics).

Summary of non-analogous characteristics and constraints for explaining Taconian tectophase differences

As mentioned there are a number of important differences between orogenic phases not yet explained by previous tectonic models. These are summarized in the following table, and are critical to development of a new model for the Taconic Orogeny.

Blountian Tectophase – Sevier Basin	Vermontian Tectophase – Martinsburg Basin
1. Iapetan basement architecture is highly rifted with narrow faults & more numerous transform segments south of Virginia Promontory; provides for less rigid rheology during collision	**1.** Iapetan basement architecture is apparently less fragmented with broader segments north of Virginia Promontory; provides for more rigid rheology during collision
2. Narrower (and shallower?) foredeep basin with sedimentation restricted to a smaller overall region	**2.** Wider foredeep basin (and deeper?) with sedimentation spreading out over a greater region
3. Foredeep architecture likely impacted by overall smaller tectonic load (accretionary wedge) spread out over a wider area with an overall lower taper angle of the load.	**3.** Foredeep architecture likely impacted by overall much larger tectonic load (accretionary wedge) constrained to an increasingly more narrow area with an overall higher taper angle of the load.
4. Widespread basal unconformity (forebulge), with deformation and accentuated erosion along GACB platform margin, developed **prior** to major pulse in sea-floor spreading (indicated by strontium isotopes); likely attributed to subduction "jamming."	**4.** Locally restricted unconformity (reactivated forebulge), with uplift and accentuated erosion along northeastern GACB platform margin (NY-PA-Adirondack/Canajoharie Arches), developed **during** major pulse in sea-floor spreading (strontium isotopes); likely developed during slab-break off leading to subduction reversal
5. Sedimentary provenance indicates minimal metamorphism, low volcanic contribution in basin fill; suggests sediments eroded from accretionary wedge derived from older passive margin sources; and presence of a forearc basin "trap" outboard of wedge that prevents these sediments from moving into foredeep basin.	**5.** Sedimentary provenance indicates increased metamorphic and volcanic contribution in foredeep basin fill; suggests sediments eroded from accretionary wedge and more heavily metamorphosed forearc to arc segments; evidently forearc basin was filled or closed tectonically.
6. Basin formed and filled more rapidly in ~5 million years (459-454 mya).	**6.** Basin formed and filled more slowly ~10 million years or more (454-444 mya) .
7. Small volcanic terranes present, but within a limited area south and north of Tennessee Embayment; First **minor** K-bentonites on platform and in foredeep after basin subsidence initiated; Most significant evidence for ash deposition late in the tectophase as basin was over-filled.	**7.** Volcanic terranes present, from Virginia promontory north; Extensive second (young) volcanic arc formed outboard of Turinian-aged terranes – not explained in previous models; **Extensive** ash deposition early in tectophase and across the platform, ash volume subsides in later filling episodes.
8. No evidence for intra-plate volcanism.	**8.** Evidence for intra-plate, through-slab extrusion of volcanics (Jonestown Volcanics & Stark's Knob Pillow Basalt); suggests a source of magma below the Laurentian plate

A NEW MODEL FOR PASSIVE MARGIN ACTIVATION: OBLIQUE COLLISION WITH TWO DISTINCT PHASES IN THE DEVELOPMENT OF THE TACONIC OROGENIC CENTER

Given the constraints discussed above, it is possible to develop a conciliatory model that helps explain the timing and spatial variations of different tectophases of the Taconic Orogeny and their impacts on the GACB and associated foreland basins. Using the physical architecture observed in the modern Banda-Timor Orogen, and an understanding of the relative rates of plate tectonic collision established in the modern (and more or less similar for the Ordovician of Laurentia based on biostratigraphy; Finney et al., 1996), it is possible to generate time series showing the relative position of key features of the foreland basin complex, and features of the hinterland. The following diagrams show ten time slice stages ranging from the pre-Knox highstand through the Ashbyan, Turinian, Chatfieldian, and into the earliest Edenian.

Stages of the Blountian tectophase and associated changes within the GACB

The first set of diagrams (**figures 14& 15**), shows the pre-Knox configuration (Tinmouth Phase of the Taconic Orogeny) through the end of the Blountian tectophase and illustrates the architectural changes of the foreland basin during stages 1 through 5 as described below.

- Stage 1): Tinmouth phase; Pre Late-Ordovician configuration of the GACB prior to the Knox Unconformity; Indian River deposited on forebulge well off the shelf margin.
- Stage 2): Uplift of the pre-Knox shelf margin (forebulge) during a global sea-level lowstand and accentuation of the Knox Unconformity; Accretionary wedge obliterates older subduction zone/trench; Subduction is "jammed."
- Stage 3): Early to mid Ashbyan; foredeep-trench complex arrival on the outer shelf-slope; Rapid growth of the accretionary prism as it encountered thicker slope and rise sediments
leading to loading, subsidence, and starvation of the Sevier foredeep; Coincident with deposition of sequences M1A and M1B on the GACB as shown on **figure 10**.
- Stage 4): Late Ashbyan to Early Turinian; Sequences M2 and M3 of **figure 10**; Onset of major flysch- filling in the Sevier Basin in the Tennessee Embayment after emergence of accretionary wedge off the Alabama-Virginia promontories; Synsedimentary deformation and destabilization of carbonate platform observed near Virginia Promontory (i.e. Fincastle Conglomerate etc.) and in foredeep of Tennessee Embayment, Foredeep just off the pre-Knox shelf edge in the Pennsylvania Embayment; Forebulge activated uplift of Tazewell and Adirondack Arches inboard of Virginia and New York promontories; Accretionary prism growth in the

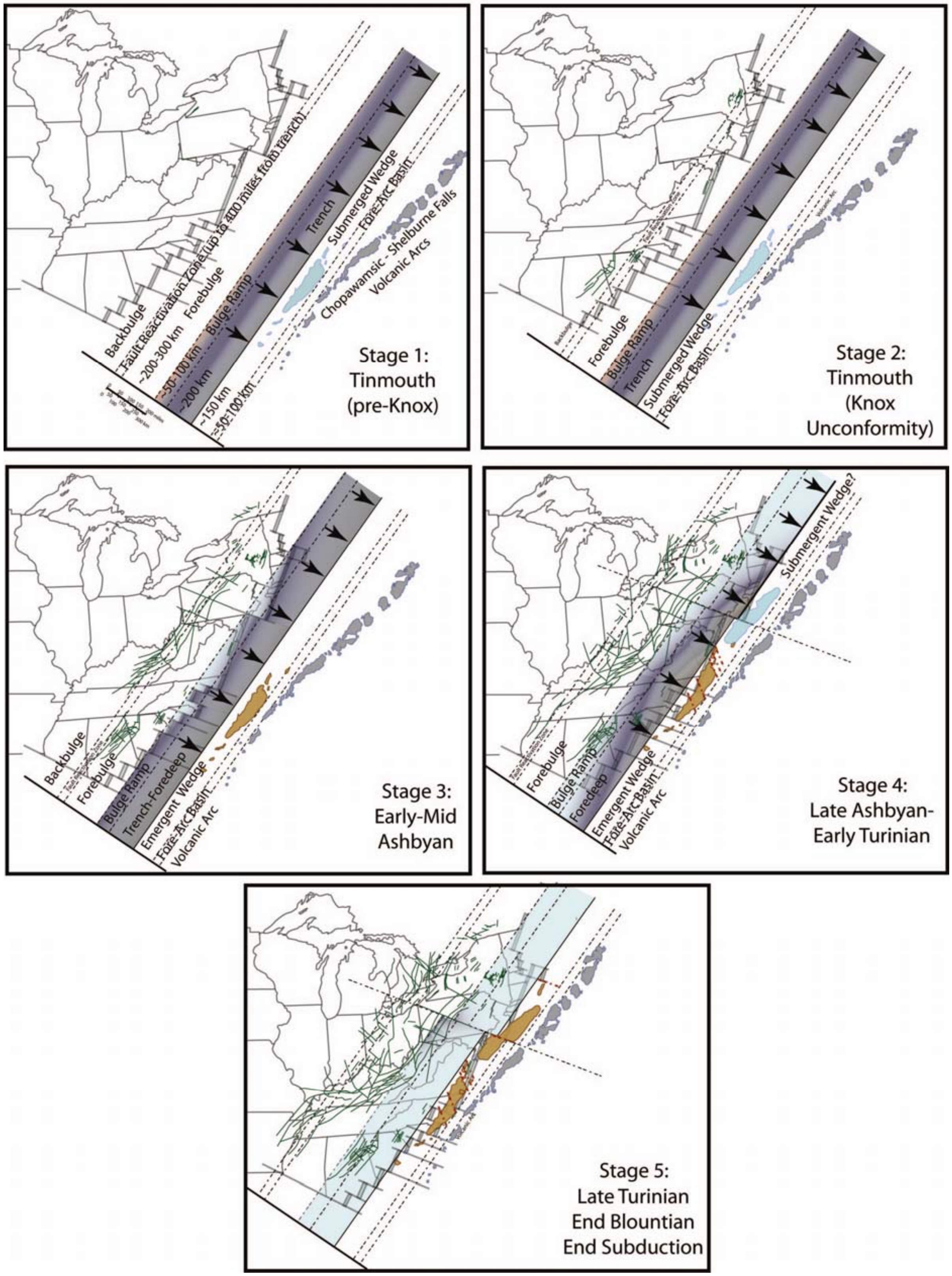


Figure 14: Time slice maps for the early Late Ordovician during the onset of plate tectonic collision. Key structural features of the foreland basin complex and hinterland terrains are positioned relative to the ancestral Laurentian margin. East-dipping subduction is also shown dominating the development of the eastern GACB through Stage 4. In contrast, no subduction zone is shown in Stage 5, i.e. end Blountian, as it is the time interval within which slab-break off occurred.

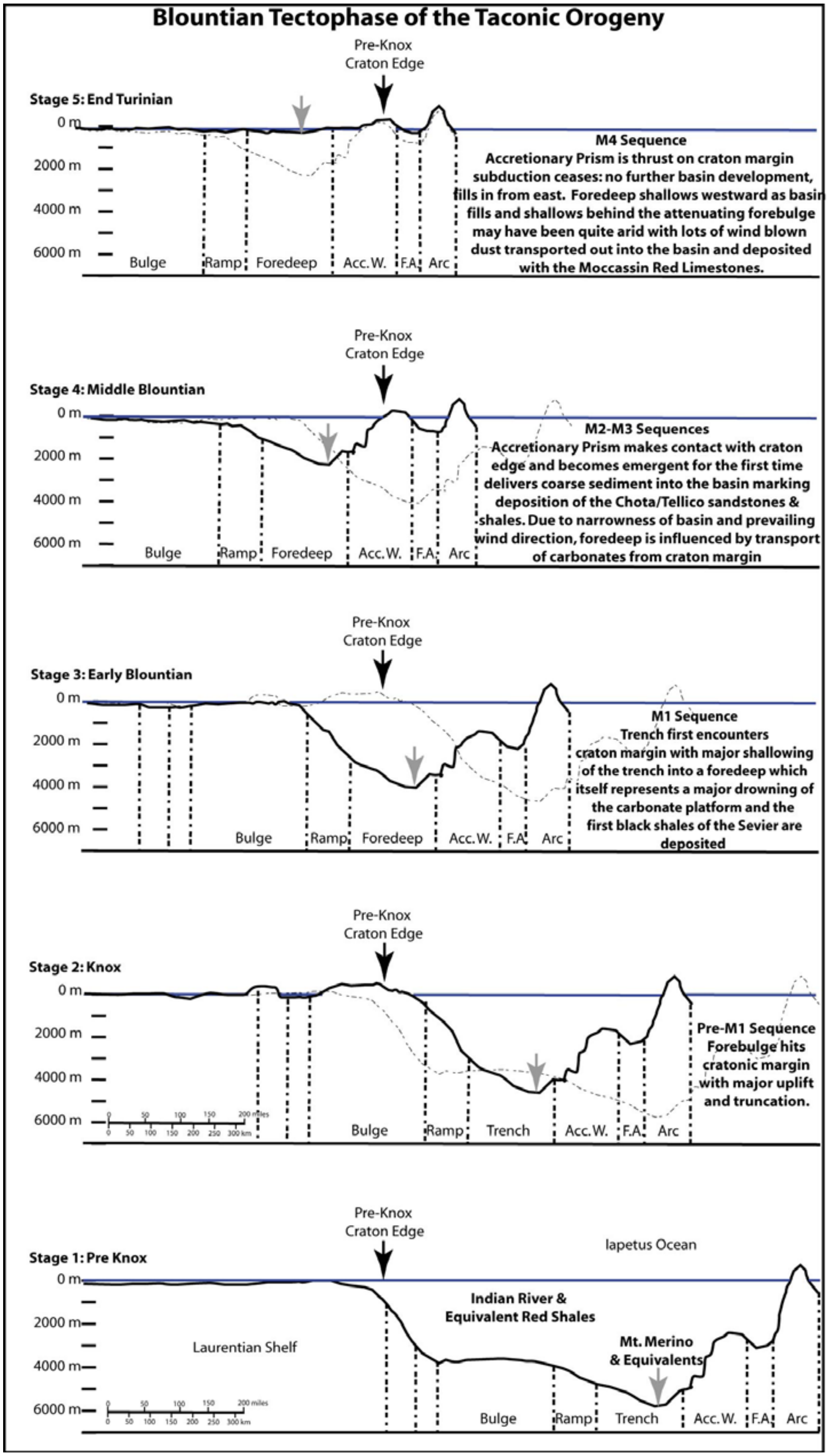


Figure 15: Time slice cross-sections for the early Late Ordovician during the Blountian tectophase of the Taconic Orogeny. To be compared with figure 14. In each diagram, the position of the former stage topographic profile is shown as a grey dotted line to show the inferred change between each stage, as is the position of the depocenter, i.e. grey arrow.

vicinity of the New York Promontory as the wedge encounters increased thicknesses of slope and rise sediments.

- Stage 5): Mid to Late Turinian; Sequences M4A-M4B of **figure 10**; Rapid shallowing across the GACB and in foredeep as recorded both in carbonate facies and via the progradation of coarser sandstones and red beds from the south; Northern extension of foredeep stalled just off the New York Promontory – Pennsylvania Embayment; Adirondack Arch is stationary, but being on-lapped by shallow-water carbonates; Only minor evidence for local faulting on portions of the Adirondack Arch suggests quiescence; Substantial evidence for explosive volcanism and increased rates of sea-floor spreading; End of Blountian tectophase

Thus, stages 1-5 (sequences M1A-M4B), record the docking of the accretionary prism with the Laurentian margin above an east-dipping subduction zone. Docking followed the short-term "jamming" of the subduction zone (during the Knox Unconformity), and resulted in the loading and depression of the Sevier Basin (foredeep) and subsidence in portions of the Champlain Trough, which was likely in the area of the back bulge basin. It also resulted in the uplift of the Beauharnois Arch, Adirondack Arch and their southern equivalents (i.e. the Tazewell Arch) within the area of the peripheral bulge formed initially during the Knox Unconformity. Synsedimentary faulting and seismite occurrences in the Sevier Basin, the abundance of intraformational limestone conglomerates, especially near uplifted arches, combined with the onset of major restriction of the GACB (in back bulge areas) initiating with siliciclastic event 3 during the Chazy-Black River transition (late M2-M3 sequences; **see Chapter 6**) collectively reflect the influence of this collision. The presence of the Adirondack Arch and its southern equivalents effectively shielded the GACB interior (i.e. the immense Black River-High Bridge lagoon) from the impact of siliciclastic sedimentation. There is evidence, however, for unique temporal windows within which siliciclastics (i.e. siliciclastic events 3, 4, and 5) were delivered, presumably during progradation of late regressive systems tracts, and/or during arid lowstand events.

The influence of the Adirondack Arch became substantially less late in the Turinian (M4A sequence) coincident with the intensification of large-scale volcanic eruptions and the appearance of molasse deposits not only in the Sevier Basin, but also in more interior areas. After the Laurentian margin failed to subduct, the Blountian tectophase ended sometime after the detachment and rapid melting of the subducted slab. Without the influence of slab-pull of the subducting plate, slab-break off initiated buoyant uplift of the former subduction zone (the foredeep) and portions of the accretionary prism just prior to the M4A sequence. This was critical because it not only stopped migration of the foreland basin complex, but it initiated tectonic-induced shallowing of the foredeep. This is signaled by the rapid progradation of accretionary wedge-derived sediments including the Walker Mountain Sandstone of the southern Appalachians and equivalents in Pennsylvania. These materials then effectively filled the foredeep from the southeast with the first molasse deposition in the M4A sequence. Likewise, during this time successions within the craton (i.e. Jessamine Dome) show the progradation of carbonate depocenters to the southeast toward the Adirondack-Tazewell Arches (as documented by Cornell, 2004; and shown schematically in **figure 10**). The infilling of the Sevier Basin, and the onlap/overlap of the Adirondack Arch essentially flattened the Blount foreland basin complex in the south and prepared the GACB for the onset of Trenton deposition at the outset of the Vermontian tectophase (sequence M5A). Farther north, in the Pennsylvania Embayment, the remnants of the Blountian foredeep was still filling, and sediments were still being trapped outboard of the shelf margin, although this changed rapidly in subsequent events.

Stages of the Vermontian tectophase and associated changes within the GACB

The second set of five time stages (**figures 16 & 17**) show the second and substantially more major basin-forming phase of the Taconic Orogeny during the Vermontian tectophase. Based on important observations, not only from the GACB, but also from the foreland as documented previously, the Vermontian tectophase is inferred to have resulted in the following changes to the architecture of the GACB.

Stage 6): Early Chatfieldian (Rocklandian); Sequence M5A; Black River-Trenton transition; Relatively gentle topography of the GACB, formed at the close of the Blountian, is inherited into the early Chatfieldian during a period of substantial sea-level rise coincident with a period of high sea-floor spreading rates (strontium excursion 2 of Chapter 6); Widespread development of clean, coarse-grained limestone facies across GACB as siliciclastic sources in the hinterland were drowned; Followed immediately by deposition of siliciclastic sediments derived from Taconic sources just after maximum flooding; First evidence for synsedimentary deformation and faulting in craton interior (400 km to the interior of the platform margin) recorded by seismites in the Jessamine Dome and by formation of the Sebree Trough during highstand; Soft-sediment deformation, and atypical (thick) turbiditic limestones deposited in central Pennsylvania, and submarine cave collapse features occur in New York, collectively indicate destabilization and inversion of topography beginning to occur in proximal edge of foredeep; Widespread distribution of siliciclastics attributed to reorganization of marine circulation patterns within the GACB in response to structural developments in the tectonic load architecture and deformation of the foreland basin complex (i.e. back bulge, forebulge, foredeep); Substantial new volcanism outboard of the Shelburne Falls Arc (i.e. the Bronson Hill Arc) east of the Turinian-aged (and older) volcanic arc system; Renewed contraction in the hinterland is inferred as a result of growth in the accretionary wedge complex producing a much larger tectonic load and renewed subsidence of foredeep; Subtle changes in sand and clay composition indicate onset of change in the load structure with minor, but increased volcanic input in sediments of the foredeep.

- Stage 7): Late-Early Chatfieldian (Kirkfieldian); Sequence M5B shown in **figure 10**; Volcanism on the new volcanic arc (Bronson Hill/Ammonoosuc Arcs) well underway; Foredeep reactivating inboard of pre-Knox shelf edge; Subsidence and drowning of southern and eastern portions of Adirondack Arch in New York and

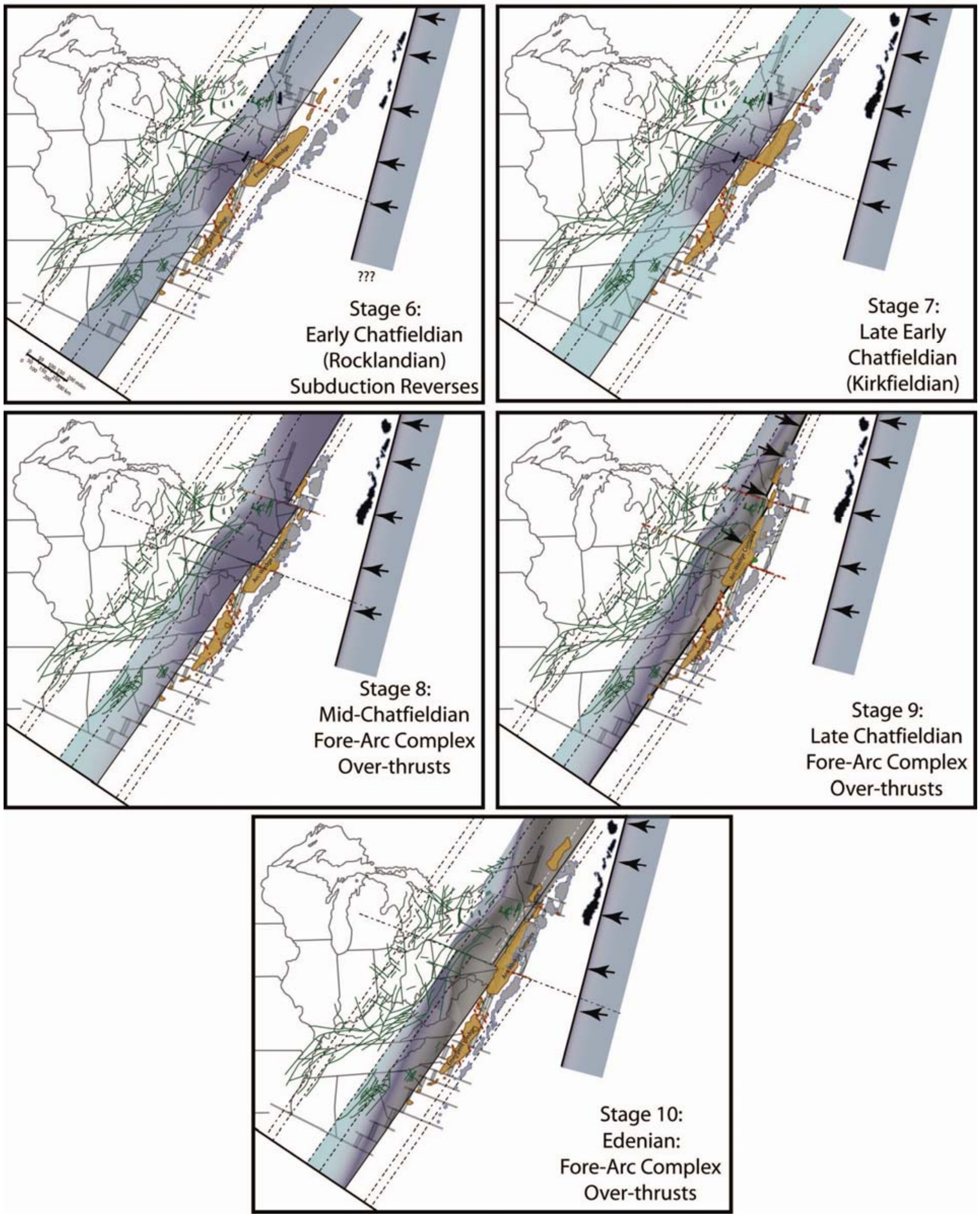


Figure 16: Time slice maps for the Late Ordovician (Chatfieldian-Edenian) during the reversal of subsidence and associated shortening (over-thrusting) of the accretionary wedge- fore arc complex. Key structural features of the Vermontian-phase foreland basin complex and hinterland terrains are positioned relative to the ancestral Laurentian margin. West-dipping subduction initiates in Stage 6, followed by significant contraction in the accretionary wedge – fore arc basin interval leading into the late Chatfieldian. Foreland deepening events coincide with major pulses in thrust events leading to growth and westward migration of the wedge complex, although at a slower rate than in the Blountian.

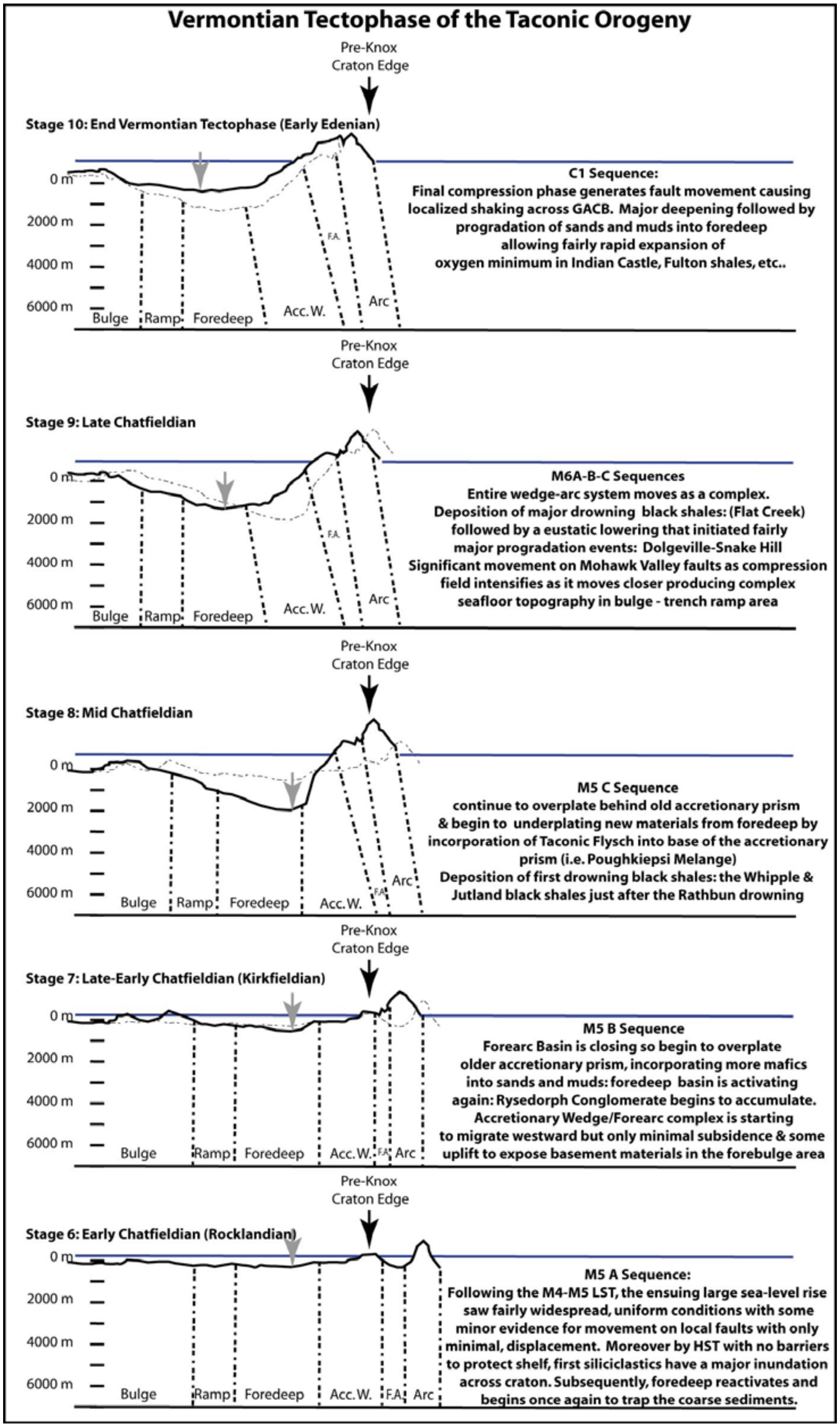


Figure 17: Time slice cross-sections for the Late Ordovician during the Vermontian tectophase of the Taconic Orogeny. To be compared with figure 16. In each diagram, the position of the former stage topographic profile is shown as a grey dotted line to show the inferred change between each stage, as is the position of the depocenter, i.e. grey arrow.

Pennsylvania initiated as limestones (Jacksonburg, Salona, and Glens Falls limestones) show deepening-upward motifs; Destabilization of slopes indicated by deposition of numerous intraformational conglomerates and breccias along newly formed forebulge-foredeep ramps (i.e. Rysedorph, Hershey, etc.); Fault-enhanced inversion of topography initiated with uplift in portions of the Adirondack Arch producing deeply incised horst blocks (even down to basement); Coarse-grained carbonate shoal complexes (i.e. Kings Falls, Hull, Kirkfield, Grier limestones, etc.) predominate craton interior between block uplift areas including Jessamine Dome, Nashville Dome, Marmora Arch, etc.; Inversion of topography (across 400+ km) suggests new period of growth in accretionary wedge complex and propagation of compressional stress inboard producing movements on ancestral faults; Provenance data show increased percentages of mafic-derived sediments indicating uplift and closure of the forearc basin well underway; Possible timing of through-slab extrusion of Stark's Knob pillow basalts in eastern New York to western Vermont.

- Stage 8): Mid-Chatfieldian (Early Shermanian); M5C-M6A Sequences of **figure 10**; Pronounced shallowing via tectonic uplift of the southern Jessamine Dome (within the forebulge area) at the end of the M5C followed by second major eustatic deepening of the Trenton in the M6A sequence accompanied again by a surge in sea-floor spreading (strontium excursion 3 of Chapter 6); Renewed period of widespread soft-sediment deformation and seismite development in Kentucky, New York, and Pennsylvania (i.e. Brannon, Russia, Milesburg limestones respectively); Deepening enhanced in foredeep of New York and Pennsylvania by depression of foredeep basin as it pulsed westward as evidenced by the first major shale deposition (i.e. Flat Creek-Canajoharie, Whipple-Stony Point, Jutland-Bush Kill, etc.); Poughkeepsie Melange of eastern New York incorporated into frontal slices of accretionary prism complex; Pronounced volcanic sedimentary signatures in clay and sandstone compositions suggesting forearc basin and volcanic arc have been incorporated into accretionary wedge as over-thrust complexes; Over-thrusts drive migration of the load west of the pre-Knox shelf margin position for the first time; possible timing of through-slab extrusion of submarine volcanics in foredeep areas of southeastern Pennsylvania (i.e. Jonestown Volcanics) if not occurring during stage 8;
- Stage 9): Late Chatfieldian (Late Shermanian), Sequences M6B-M6C shown in **figure 10**; Following rapid growth of the accretionary prism complex in Stage 8, this stage shows continued black shale deposition in the foredeep of New York and Pennsylvania (Valley Brook/Upper Flat Creek, Iberville, Ramseyburg, etc.), but

these are accompanied by pronounced progradation of coarser siliciclastics from the east and south (i.e. Snake Hill of New York, Bangor Beds (middle Martinsburg) of eastern Pennsylvania, and Strodes Creek-Millersburg of Kentucky); Likewise carbonates also show pronounced shallowing and progradation into the foredeep and cratonic interior troughs leading into the C1 sequence; Progradation was accompanied by another phase of intense synsedimentary deformation from Kentucky to New York suggesting some evidence for faulting leading into the last sea-level lowstand of the Chatfieldian (i.e. late Trenton Group).

- Stage 10): Early Edenian, Sequence C1 shown in **figure 10**; Pluton emplacement, and felsic volcanism occurring in the Bronson Hill Arc (**see figure 6**); Metamorphism and intrusions observed in Maryland-Virginia terranes (**see figure 9**) inboard of new volcanic arcs to the northeast); Last major pulse in the contraction of the Taconic orogen in eastern Pennsylvania and north of the New York Promontory; Propagation of the westernmost Taconic thrusts and Hamburg klippen into eastern New York and southeastern Pennsylvania to their westernmost position (i.e. "Logan's Line") marking the transition to the Hudson Valley phase; Most significant eustatic deepening of the Late Ordovician (strontium excursion 4) or at least most significant expansion of oxygen minimum zone possibly signaling a return to greenhouse conditions; Taconic foredeep moved to its westernmost position in the west-central Mohawk Valley in advance of the wedge; Pronounced widespread shale deposition during maximum flooding of C1 sequence (i.e. Fulton-Kope of Kentucky and Ohio, Indian Castle of New York; Antes-Pen Argyl of Pennsylvania); Taconic/Martinsburg foreland basin sees second pulse in progradation of much coarser siliciclastics late in the C1 sequence (i.e. Garrard Siltstone of Kentucky; Hasenclever-Schenectady of New York, Reedsville-New Tripoli-Shochary of Pennsylvania); GACB platform fully dominated by siliciclastics, carbonate deposition predominant only during limited windows (i.e. small scale sea-level rise events) when siliciclastic sediment supply is reduced.

Thus, the second phase of tectonic activity in the Taconic Orogeny, rather than reflecting yet another "analogous" basin-forming event, is interpreted herein to represent the reorganization of the plate margin via establishment of a new west-dipping subduction zone. During the five stages of this phase, basin subsidence was accommodated by contraction of the accretionary wedge– forearc basin– volcanic arc complex and movement of the complex inboard of the pre-

Knox continental margin. This is in contrast to the Blountian tectophase that formed due to initial docking of only the accretionary wedge with the pre-Knox continental margin followed by slab break-off. In the Vermontian tectophase, contraction of the accretionary wedge complex propagated from east to west resulting in the shortening of the tectonic load producing a much higher taper angle and as a consequence producing a much more substantial foredeep (**figure 18**). Multiple contraction events grew the wedge complex and pushed the entire complex

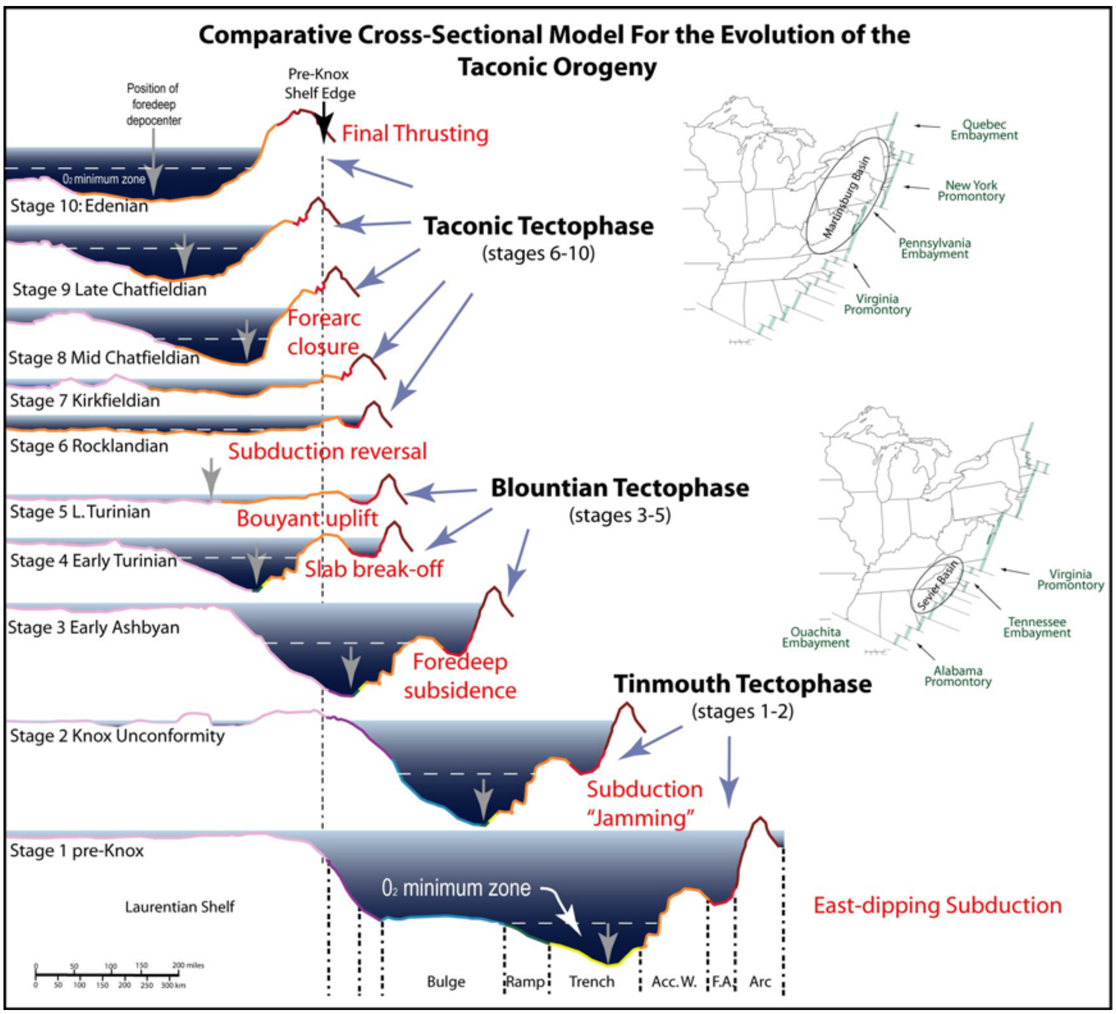


Figure 18: Summary diagram showing the "non-analogous" evolution of the Taconic Orogeny as proposed in this study. The first foreland-basin producing phase (Stages 1-5, inclusive of the Tinmouth-Blountian tectophases) records docking of the accretionary prism, followed by subduction failure and rebound/rapid filling of the foredeep. The second foreland-basin producing phases (Stages 6-10) record reorganization in subduction orientation, growth and contraction of the orogen, and migration of the orogenic load some distance (50-100 km) west of the ancestral pre-Knox shelf edge.

cratonward of the pre-Knox shelf margin to the position of eastern New York State. This structural activity in the hinterland resulted in propagation of collisional stress into the craton producing reactivation and inversion of older fault structures across the foreland basin complex (i.e. forebulge/back bulge/foredeep regions), an area of greater than 400,000 km^2. This fault activity produced a much more highly-dissected GACB platform that was prone to substantially different circulation patterns compared to any previous event. As a result, siliciclastic sediments were periodically introduced to the platform during sea-level oscillations, and effectively led to the last stand of the great American carbonate bank at the end of the C1 sequence transgression (represented by the Steuben Limestone of New York, the Point Pleasant Formation of Kentucky, and the Coleville Member of the Coburn Formation of Pennsylvania). The loss of carbonate-dominated platforms in the foreland basin and neighboring portions of the craton was clearly linked to sea level change, expansion in the oxygen minimum zone perhaps as a result of climate change. Given the emplacement of plutons, felsic volcanism, and increased rates of seafloor spreading between 450-449 mya (during strontium excursion four), it is possible that greenhouse gas emission was substantial enough to have produced greenhouse conditions following a somewhat cooler period as argued elsewhere (i.e. see Holland & Patzkowsky, 1996). Moreover the fate of widespread carbonate production in the GACB was effectively sealed thereafter by the rapid progradation of coarser-grained siliciclastics as the Taconic/Martinsburg basin filled.

IMPLICATIONS FOR LATE ORDOVICIAN PALEOCLIMATE

Through the use of a highly resolved chronostratigraphic model, afforded by the construction of detailed depositional sequences for the Late Ordovician GACB, it is possible to link environmental changes recorded in the craton with events that occurred within the hinterland. The latter of which may not have been recorded anywhere else owing to the complexity of the stratigraphic record of these regions and/or substantial alteration of these regions by metamorphism and volcanism. As such, this study has shown a possible correlation between tectonic events (i.e. changes in the rate of sea floor spreading, uplift of hinterland complexes, etc.) to pulses in subsidence and the delivery of siliciclastic sediments into foreland basin systems as well as additional changes in more interior portions of the GACB. In addition, it is possible to link these patterns to local, regional, and possibly even global patterns of sea-level change.

In this study, at least four strontium isotopic excursions provide evidence for increased rates of sea-floor spreading that are followed by rapid dumping of continent-derived sediments. It is implied herein that three of these episodes are intimately linked to phases in the growth (uplift) of accretionary wedge complexes and the pronounced subsidence of near-wedge foredeep basins. In addition, where these increased sea floor spreading events are coincident with major transgressions across the entire platform, there is strong support for increased rates of seafloor spreading to be a driving mechanism for eustatic sea-level change regardless of tectonic-induced subsidence or uplift. Thus, although each major deepening and associated black shale advance (i.e. M1B-M2, M5A, M6A, C1 sequences) across the foredeep and marginal basins may represent subsidence following the growth of successive thrust sheets in the accretionary wedge; it is likely that sea-level is also high due to pronounced displacement of seawater owing to rapid spreading. Moreover, given high sea-floor spreading rates and increased evidence for volcanism, these intervals of time also see increased degassing of CO_2 and other gases to the atmosphere. Therefore greenhouse conditions would be more pervasive and would promote the expansion of oxygen restriction in marine settings. This would promote the preservation of organic matter in the deep sea and poorly circulating marginal basins and help explain darker organic-rich shales deposited at these times.

Interestingly, following these dark shale events, depositional sequences show stepwise shallowing (i.e. M3, M4; after the M1A-M2 sequence deepening; M5B, M5C; after the M5A deepening; M6B, M6C after the M6A deepening; and C2, etc., after the C1 deepening). Given the pronounced evidence for progradation of sediments immediately following these deepening episodes, and the consumption of CO_2 during weathering reactions, the correlation can be at least suggested for a link between sea-level drop and climate change. That is, the removal of atmospheric CO_2 through weathering reactions in pre-vegetated landscapes could contribute to less intense greenhouse periods (and/or more intense icehouse phases; (i.e. see Kump & Arthur, 1999; Kump et al., 1999), slightly lowered sea-levels, and more intense ocean circulation that would continue until the ensuing pulse of sea floor spreading slowed currents and resulted in more intense stratification, and expansion of oxygen minimum zones. This linkage obviously has implications for previous models that support upwelling within the Laurentian craton, especially through the Sebree Trough as discussed in Chapter 6.

Paleogeographic reconstructions show that much of the GACB region was located within the sub-tropical high pressure belt located about 30° degrees south of the Ordovician equator (**figure 19**). Detailed lithostratigraphic and event stratigraphic studies of key regions in this study (Chapters 3, 4, 5, and 6), show sedimentologic evidence for relatively warm, arid conditions in this region during deposition of the Chazy and Black River Groups. Many of the unique time-restricted facies identified and described in Chapter 6 may have been in fact strongly tied to periods of relatively more intense aridity especially during pronounced regressive phases. With a predominance of hot dry air descending near the sub-tropical high pressure belt, surface winds would have been produced in the vicinity of the southeastern Laurentian margin and the newly uplifted accretionary wedge may have acted to intensify warming. Winds would have

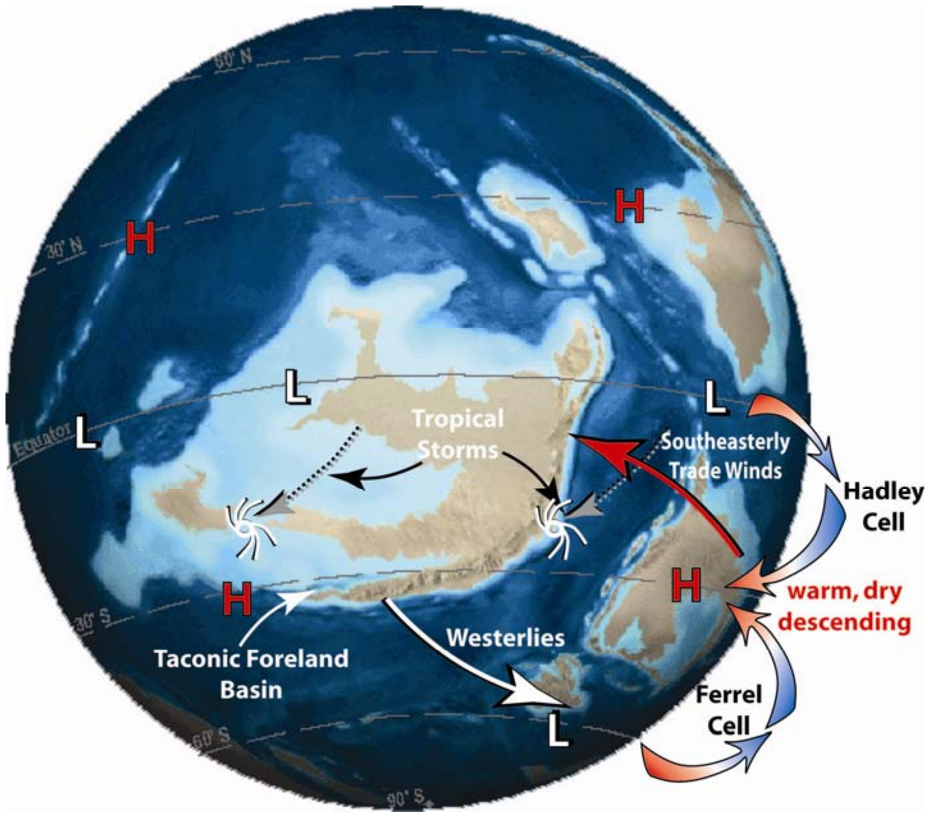


Figure 19: Paleogeographic reconstruction (modified after Blakey, 2007) and inferred atmospheric circulation patterns for Laurentia during the Late Ordovician (late Vermontian tectophase). The Taconic foreland basin is positioned near the sub-tropical high pressure belt located near 30° S with the majority of the GACB located farther north within the Hadley cell and within the limits of a sub-tropical arid belt. Surface winds would generally flow from southeast to northwest. Major tropical storms originating near the equatorial low pressure belts would be deflected toward the southwest.

then blown northwestward toward the equator. Thus predominant siliciclastic events found during latest highstand (falling-stage) to early lowstand events on the carbonate platform during the latest Ashbyan to Turinian (see chapter 6) may have been transported onto the platform as wind-blown sediments derived from the newly uplifted accretionary wedge complex.

Following the Blountian aridity events, during the Vermontian the pronounced growth in the accretionary wedge complex may have modified circulation enough to change the climatic regime of the area. The presence of a significantly more elevated terrane near the convergence of major atmospheric circulation cells (i.e. Hadley and Ferrel cells) has the potential for obstructing the flow of subtropical trade winds. At this latitude, the development of an elevated terrain has the potential to produce a significant orographic effect and/or may have initiated monsoonal climates and increased runoff in the formerly arid GACB region. Moreover, these highlands may have funneled large tropical storms, originating in more equatorial areas, along the length of the orogen resulting in the setup of strong northwest directed (shore perpendicular) storm currents. If the ultimate growth of the wedge-arc complex was completed in late Chatfieldian to Edenian time, this may help explain the predominance of north-northwest directed sediment transport as recorded by numerous gutter marks and channel features in Cincinnatian-aged strata.

SUMMARY OF THE DEMISE OF THE GREAT AMERICAN CARBONATE BANK

In order to understand the complexity of the Taconic Orogeny in eastern Laurentia and its ultimate impact on the GACB, aspects of timing, rates and vectors of development in the foreland basins produced during this orogenic event have been considered herein using the depositional sequences first outlined by Brett and colleagues (2004), and expanded on in chapter 7. Moreover, the outcome of theoretical modeling has made several predictions regarding the expected behavior of the foreland basins during the development of hinterlands during orogenesis. These theoretical models have been applied to case studies of both the Blountian and Vermontian tectophases of the orogeny in order to explain major features and patterns in the evolution of these foreland basin(s) and have attempted to establish the connections between them for the first time. While these studies have elucidated many sedimentologic and stratigraphic patterns that fit within the constraints of these early models, often the processes that are invoked to account for these sedimentologic and stratigraphic changes have poor chronostratigraphic resolution, which has ultimately led to misleading interpretations.

As such, these former models fail to adequately evaluate some key questions regarding the exact timing (and lateral extent) of local, regional, and basin-wide events, much less document the extent to which these events have impacted the shallower portions of the craton. These studies do not adequately evaluate spatial-temporal scales of each of the tectophases (Blountian, Vermontian) and their relationships to one another. Previous tectonic models have suggested that these tectophases were the result of completely separate, but analogous, collisional events related to the accretion of separate terranes to eastern Laurentia. Yet these studies fail to address even the basic observation concerning the scale of these orogenic phases and the size of the foreland basins that they produced. In addition, many predictions made by these models have not, as yet been born out in the stratigraphy of the foreland succession.

In the light of more recent theoretical modeling on foreland basin evolution, new provenance studies of the Sevier and Martinsburg (Blountian and Vermontian respectively) sedimentary basins, and a refined high-resolution chronostratigraphic framework for eastern Laurentia, presented herein, the understanding of the Taconic Orogeny and its impact on the foreland have been improved and updated to a more parsimonious model. Through the development of a sequence stratigraphy-based chronostratigraphic framework from the craton into the foredeep(s) of the Taconic Orogeny, it is now possible to answer outstanding questions regarding the development and migration of foredeep-forebulge-back bulge regions of the foreland. Moreover, this analysis affords an increased understanding of the relative timing, position, and scale of each of these tectophases, and a better understanding of events occurring in the hinterland as recorded by their sedimentary response.

Salient outcomes of this study include:

- The construction of a new model for the Taconic Orogeny, building on and integrating older models. Through the synthesis of a multitude of data accumulated in the quarter decade since the last tectonic models were produced, an evolved view of Late Ordovician orogenesis is proposed here.

- As suggested above, the two primary phases of the Taconic Orogeny are now considered to represent distinct phases of the same protracted collisional event (as opposed to analogous collisional events) separated in time by a major plate tectonic reorganization episode (i.e. subduction reversal).
- Spatial-temporal constraints, based on the size and duration of each basin-forming episode, are linked to prominent changes in the architecture of the hinterland during contraction and telescoping of stacked thrust-sheets in the hinterland after subduction reversal occurred.
- The timing of intense volcanic ash deposition in the late Turinian-early Chatfieldian, and the occurrence of enigmatic through-slab extrusion of volcanics slightly later, is now integrated and related to rapid melting of the detached slab and the evolution of magmas in the vicinity of the newly formed west-dipping plate located below the hinterland of the former subduction zone.
- A plausible correlation is tentatively proposed linking multiple pulses of sea-level change recorded in strata across the GACB with both tectonic loading/basin subsidence episodes as well as changes in the rates of seafloor spreading.

This study has implications not only for further studies in the Late Ordovician basins of eastern Laurentia, but in the understanding of the behavior of similar foreland basin systems and indeed the modern Timor-Banda orogen. Is it possible that the failure of subduction off the northern coast of Australia today, and the recently noted movement on the Wetar and Flores thrusts north of the Banda Arc, signify an end to Blountian-style foreland basin development in that region? Will the future see: rapid infilling of the Timor Trough and modification of oceanic currents, large-scale volcanic eruptions on the scale of the Late Ordovician Millbrig event, a subduction reversal as is presently postulated?

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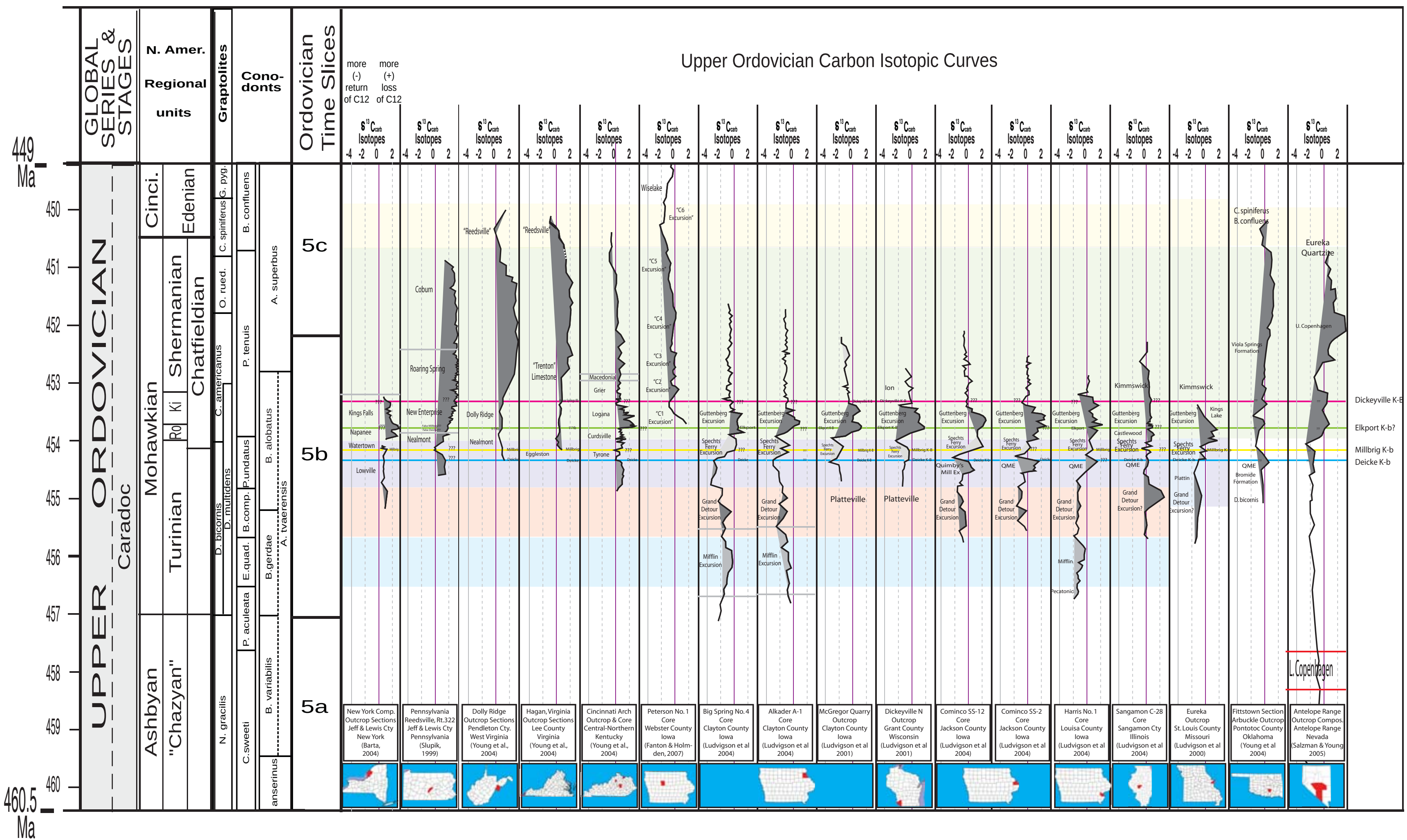
Appendix A1:

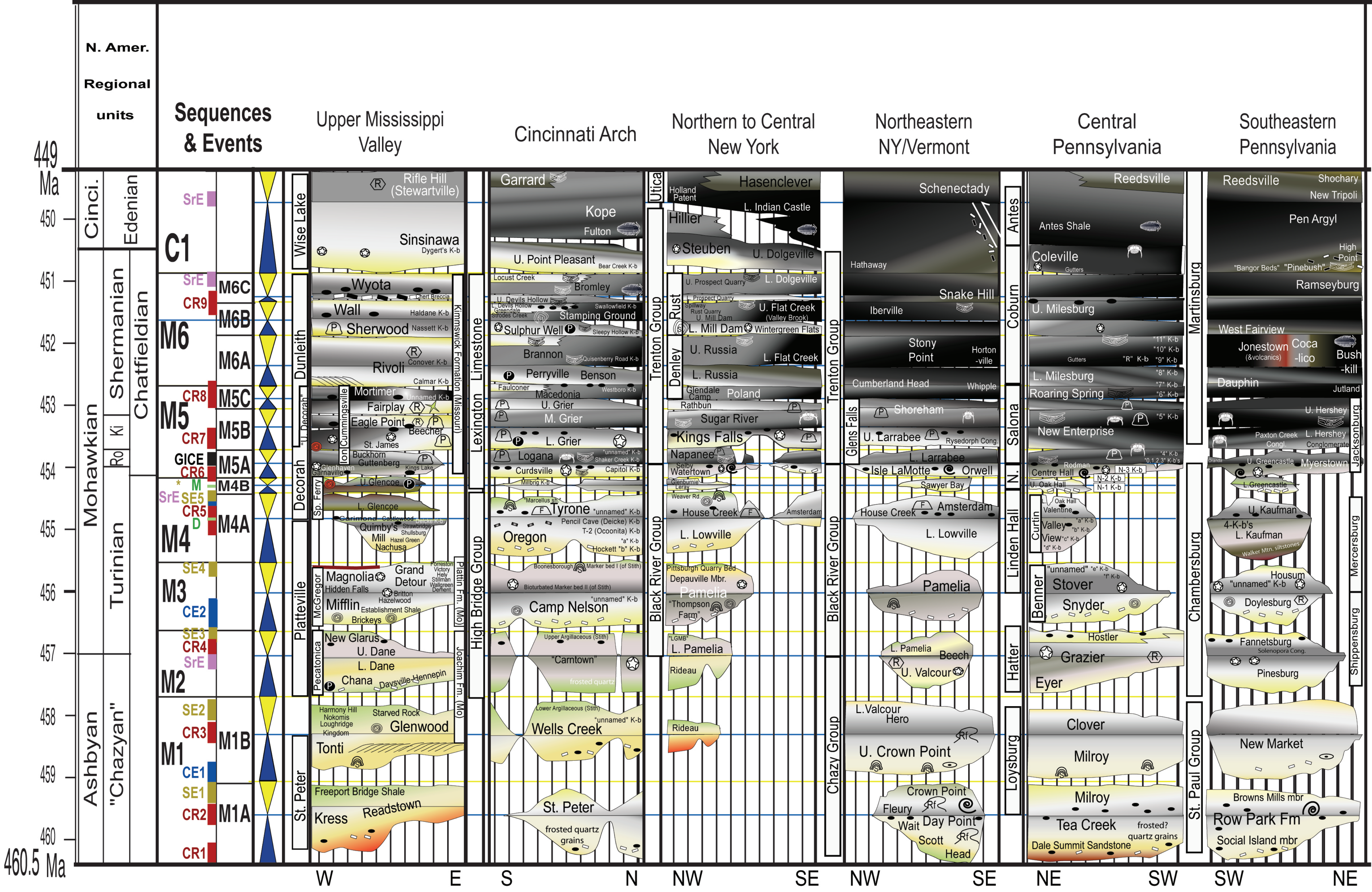
Chapter 2, Figure 9: Carbon isotopic curves for localities across the GACB region including New York, Pennsylvania, West Virginia, Virginia, Iowa, Wisconsin, Illinois, Missouri, and Oklahoma. Also shown is the profile for Nevada. The stratigraphic position of profiles is established relative to the position of the four K-bentonites: Deicke, Millbrig, Elkport, and Dickeyville as currently established. Isotopic curves modified from data presented by: Barta (2004), Fanton & Holmden (2007), Ludvigson et al., (2000, 2001, 2004), Salzman & Young, (2005), Slupik (1999), Young et al., (2004)



Upper Ordovician Carbon Isotopic Curves
GLOBAL SERIES & STAGES
N. Amer. Regional units
Graptolites
Conodonts
Ordovician Time Slices
more (-) return of C12
more (+) loss of C12
UPPER ORDOVICIAN
Caradoc
Mohawkian
Cinci.
Edenian
Shermanian
Chatfieldian
Turinian
Ashbyan
"Chazyan"
5c
5b
5a
449 Ma
460.5 Ma
Dickeyville K-B
Elkport K-b?
Millbrig K-b
Deicke K-b
New York Comp. Outcrop Sections Jeff & Lewis Cty New York (Barta, 2004)
Pennsylvania Reedsville, Rt.322 Jeff & Lewis Cty Pennsylvania (Slupik, 1999)
Dolly Ridge Outcrop Sections Pendleton Cty. West Virginia (Young et al., 2004)
Hagan, Virginia Outcrop Sections Lee County Virginia (Young et al., 2004)
Cincinnati Arch Outcrop & Core Central-Northern Kentucky (Young et al., 2004)
Peterson No. 1 Core Webster County Iowa (Fanton & Holmden, 2007)
Big Spring No. 4 Core Clayton County Iowa (Ludvigson et al 2004)
Alkader A-1 Core Clayton County Iowa (Ludvigson et al 2004)
McGregor Quarry Outcrop Clayton County Iowa (Ludvigson et al 2001)
Dickeyville N Outcrop Grant County Wisconsin (Ludvigson et al 2001)
Cominco SS-12 Core Jackson County Iowa (Ludvigson et al 2004)
Cominco SS-2 Core Jackson County Iowa (Ludvigson et al 2004)
Harris No. 1 Core Louisa County Iowa (Ludvigson et al 2004)
Sangamon C-28 Core Sangamon Cty Illinois (Ludvigson et al 2004)
Eureka Outcrop St. Louis County Missouri (Ludvigson et al 2000)
Fittstown Section Arbuckle Outcrop Pontotoc County Oklahoma (Young et al 2004)
Antelope Range Outcrop Compos. Antelope Range Nevada (Salzman & Young 2005)

Appendix A2:

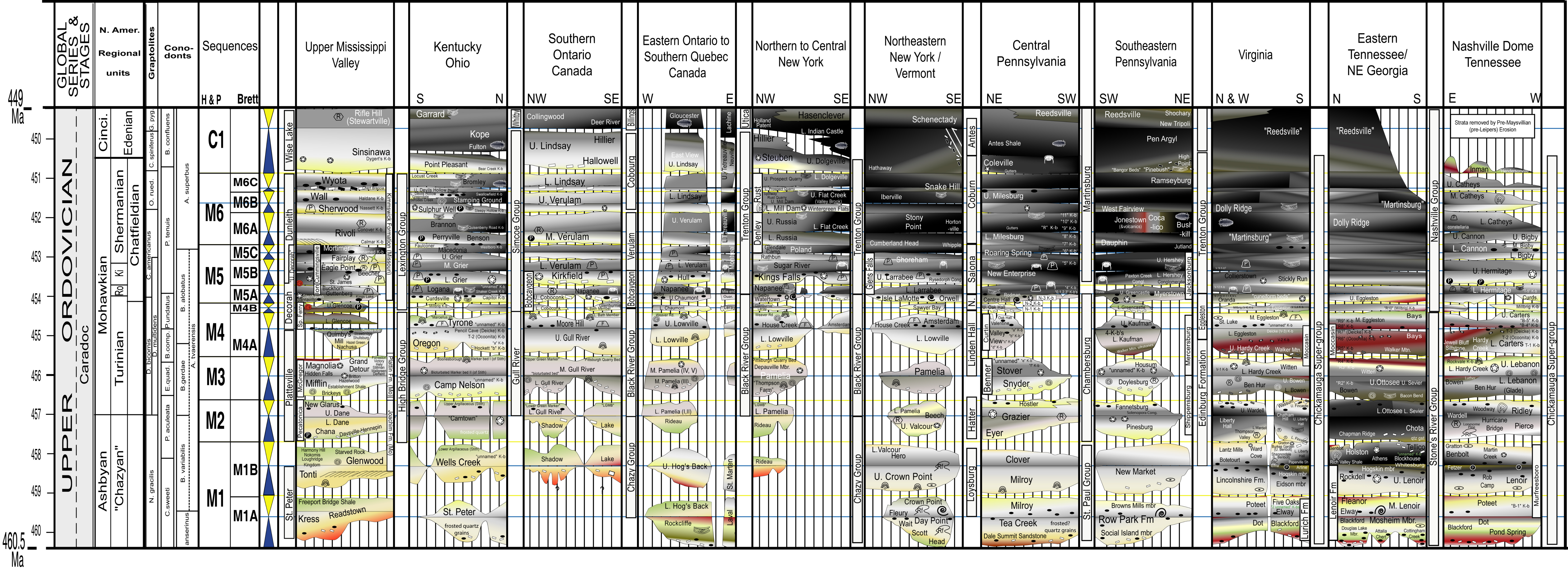
Chapter 7, Figure 1: Depositional sequences for the Cincinnati Arch, New York-Ontario Platform, and central and eastern Ridge and Valley Province of Pennsylvania. Also shown are key biostratigraphic zones and events recognized through detailed investigations discussed elsewhere. Event abbreviations: CR1: Chert-Rich Interval 1, SE1: Siliciclastic Event 1, CE1: Calcification Event 1, SrE: Strontium Isotopic Excursion, M: Millbrig K-bentonite, D: Deicke K-bentonite, GICE: Guttenburg Isotopic Carbon Excursion..



N. Amer. Regional units
Sequences & Events
Upper Mississippi Valley
Cincinnati Arch
Northern to Central New York
Northeastern NY/Vermont
Central Pennsylvania
Southeastern Pennsylvania
449 Ma
450
451
452
453
454
455
456
457
458
459
460
460.5 Ma
Cinci.
Edenian
Mohawkian
Shermanian
Chatfieldian
Ki
Ro
Turinian
Ashbyan
"Chazyan"
C1
M6
M5
M4
M3
M2
M1
SrE
CR9
CR8
CR7
GICE
CR6
M
SE5
CR5
D
SE4
CE2
SE3
CR4
SE2
CR3
CE1
SE1
CR2
CR1
M6C
M6B
M6A
M5C
M5B
M5A
M4B
M4A
M1B
M1A
Wise Lake
Dunleith
Decorah
"U. Decorah"
Sp. Ferry
Platteville
McGregor
Pecatonica
St. Peter
Cummingsville
Kimmswick Formation (Missouri)
Plattin Fm. (Mo)
Joachim Fm. (Mo)
Rifle Hill (Stewartville)
Sinsinawa
Dygert's K-b
Wyota
Chert Breccia
Wall
Haldane K-b
Sherwood
Nassett K-b
Rivoli
Conover K-b
Calmar K-b
Mortimer
Unnamed K-b
Fairplay
Eagle Point
Beecher
St. James
Buckhorn
Guttenberg
Glenhaven
Garnavillo
Kings Lake
U. Glencoe
L. Glencoe
Quimby's Mill
Nachusa
Carimona
Castlewood
Strawbridge
Shullsburg
Hazel Green
Magnolia
Hidden Falls
Grand Detour
Forreston
Victory
Hely
Stillman
Wallgreen
Dement
Mifflin
Britton
Hazelwood
Establishment Shale
Brickeys
New Glarus
U. Dane
L. Dane
Chana
Daysville-Hennepin
Harmony Hill
Nokomis
Loughridge
Kingdom
Starved Rock
Glenwood
Tonti
Freeport Bridge Shale
Kress
Readstown
Lexington Limestone
High Bridge Group
Garrard
Kope
Fulton
U. Point Pleasant
Bear Creek K-b
Locust Creek
Bromley
U. Devils Hollow
L. Devils Hollow
Greendale
Strodes Creek
Swallowfield K-b
Stamping Ground
Sulphur Well
Sleepy Hollow K-b
Brannon
Quisenberry Road K-b
Perryville
Benson
Faulconer
Westboro K-b
Macedonia
U. Grier
M. Grier
L. Grier
Logana
"unnamed" K-b
Shaker Creek K-b
Curdsville
Capitol K-b
Millbrig K-b
"Marcellus sh"
Tyrone
"unnamed" K-b
Pencil Cave (Deicke) K-b
T-2 (Oconita) K-b
"a" K-b
Oregon
Hockett "b" K-b
Boonesborough
Marker bed I (of Stith)
Bioturbated Marker bed II (of Stith)
"unnamed" K-b
Camp Nelson
Upper Argillaceous (Stith)
"Camtown"
frosted quartz
Lower Argillaceous (Stith)
"unnamed" K-b
Wells Creek
St. Peter
frosted quartz grains
Utica
Trenton Group
Rust
Denley
Black River Group
Holland Patent
Hasenclever
L. Indian Castle
Hillier
Steuben
U. Dolgeville
U. Prospect Quarry
L. Dolgeville
L. Prospect Quarry
Spillway
Rust Quarry
U. Mill Dam
U. Flat Creek
(Valley Brook)
L. Mill Dam
Wintergreen Flats
U. Russia
L. Flat Creek
L. Russia
Glendale
Camp
Poland
Rathbun
Sugar River
Kings Falls
Napanee
Selby
Watertown
Glenburnie
Leray
Weaver Rd.
House Creek
Amsterdam
L. Lowville
Pittsburg Quarry Bed
Depauville Mbr.
Pamelia
"Thompson Farm"
"LGMB"
L. Pamelia
Rideau
Rideau
Trenton Group
Black River Group
Chazy Group
Glens Falls
Schenectady
Hathaway
Snake Hill
Iberville
Stony Point
Horton -ville
Cumberland Head
Whipple
Shoreham
U. Larrabee
Rysedorph Cong.
L. Larrabee
Isle LaMotte
Orwell
Sawyer Bay
House Creek
Amsterdam
L. Lowville
Pamelia
L. Pamelia
Beech
U. Valcour
L.Valcour
Hero
U. Crown Point
Crown Point
Fleury
Wait
Day Point
Scott
Head
Antes
Coburn
Salona
N.
Linden Hall
Curtin
Benner
Hatter
Loysburg
Reedsville
Antes Shale
Coleville
Gutters
U. Milesburg
"11" K-b
"10" K-b
"9" K-b
Gutters
"R" K-b
L. Milesburg
"8" K-b
"7" K-b
"6" K-b
Roaring Spring
"5" K-b
New Enterprise
"4" K-b
"0,1,2,3" K-b's
Rodman
Centre Hall
N-3 K-b
N-2 K-b
N-1 K-b
U. Oak Hall
L. Oak Hall
Valentine
Valley View
"a" K-b
"b" K-b
"c" K-b
"d" K-b
"unnamed" "e" K-b
"f" K-b
Stover
Snyder
Hostler
Grazier
Eyer
Clover
Milroy
Milroy
Tea Creek
frosted? quartz grains
Dale Summit Sandstone
Martinsburg
Chambersburg
St. Paul Group
Jacksonburg
Mercersburg
Shippensburg
Reedsville
Shochary
New Tripoli
Pen Argyl
High Point
"Bangor Beds"
"Pinebush"
Ramseyburg
West Fairview
Jonestown
(&volcanics)
Coca -lico
Bush -kill
Dauphin
Jutland
U. Hershey
Paxton Creek Congl.
L. Hershey Conglomerate
Granda
U. Greencastle
Myerstown
L.Greencastle
U. Kaufman
4-K-b's
L. Kaufman
Walker Mtn. siltstones
Housum
"unnamed" K-b
Doylesburg
Fannetsburg
Solenopora Cong.
Pinesburg
New Market
Browns Mills mbr
Row Park Fm
Social Island mbr
W
E
S
N
NW
SE
NW
SE
NE
SW
SW
NE

Appendix A3:

Depositional sequences of the Cincinnati Arch, New York-Ontario Platform, and central and eastern Ridge and Valley Province of Pennsylvania shown in relation to inferred depositional sequences elsewhere in eastern North America including the Nashville Dome, Eastern Tennessee, Virginia, and the Upper Mississippi Valley. Depositional sequences from Virginia and Tennessee reflect a modification of the sequence framework of Holland & Patzkowsky (1996, 1998).


GLOBAL SERIES & STAGES
N. Amer. Regional units
Graptolites
Conodonts
Sequences
H & P
Brett
Upper Mississippi Valley
Kentucky Ohio
Southern Ontario Canada
Eastern Ontario to Southern Quebec Canada
Northern to Central New York
Northeastern New York / Vermont
Central Pennsylvania
Southeastern Pennsylvania
Virginia
Eastern Tennessee/ NE Georgia
Nashville Dome Tennessee
449 Ma
460.5 Ma
UPPER ORDOVICIAN
Caradoc
Mohawkian
Cinci.
Edenian
Shermanian
Chatfieldian
Turinian
Ashbyan
"Chazyan"
C1
M6
M6C
M6B
M6A
M5
M5C
M5B
M5A
M5B
M4
M4A
M3
M2
M1
M1B
M1A
Strata removed by Pre-Maysvillian (pre-Leipers) Erosion
Chickamauga Super-group
Nashville Group
Stones River Group
Trenton Group
Black River Group
Chazy Group
Lexington Group
High Bridge Group
Simcoe Group
Gull River
Martinsburg
Chambersburg
St. Paul Group
Edinburg Formation
Reedsville
Kope
Sinsinawa
Garrard
Collingwood
Hasenclever
Schenectady
Camp Nelson
Tyrone
Oregon
Wells Creek
St. Peter
Kings Falls
Lowville
Pamelia
Crown Point
Grazier
Clover
Milroy
Tea Creek
Dolly Ridge
Row Park Fm
New Market
Lincolnshire Fm.
Carters
Lebanon
Ridley
Pierce
Murfreesboro
Pond Spring